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UNION GÉODÉSIQUE ET GÉOPHYSIQUE INTERNATIONALE

Association Internationale
d'Hydrologie Scientifique

ASSEMBLÉE GÉNÉRALE D'OSLO
19-28 août 1948

TOME III

Travaux de la Commission des Eaux Souterraines

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LOUVAIN (Belgique)

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COMMISSION DES EAUX SOUTERRAINES

COMPTES-RENDUS DES SÉANCES

Séance du 23 août 1948

La séance est ouverte à 9 h 45 par M. Slettenmark qui annonce que le Comité exécutif de L'Association a décidé de donner la présidence de la Commission à M. A. Nelson Sayre pour remplacer M. Meinzer décédé.

Le Bureau des séances est donc constitué par M. Sayre, président et M. Frolow, secrétaire.

M. Sayre prononce un éloge ému de O. E. Meinzer et de son œuvre dont les résultats sont importants dans la vie et le développement des Etats-Unis.

La parole est ensuite donnée à M. Renqvist pour l'exposé du rapport de M. Matti Wäre. A la suite de cet exposé M. Renqvist précise, sur la demande de M. Frolow, que les échantillons ont été prélevés sous forme de monolithes de 70 cm.

Le Dr. Deij a ensuite la parole pour exposer son rapport sur les lysimètres.

M. Sayre, président, félicite le rapporteur et lui demande de préciser 1) les moyens employés pour protéger les lysimètres des eaux de ruissellement et 2) le prix de l'installation.

M. Deij explique que le ruissellement n'existe pas autour de l'installation lysimétrique et que le prix de l'installation s'est élevé à un million de guldens.

M. Vibert pense que l'équation du bilan donné par M. Deij est bien valable pour le lysimètre, mais qu'elle peut ne pas l'être pour le terrain en place. M. Deij répond que dans les conditions de son installation il ne doit pas y avoir de différence.

M. G. W. Musgrave said : The author has made two points on which I heartly agree. Lysimeters as instruments of study of hydrologic problems should not be discarded. Despite criticisms of them that have been made in times past, they may serve a very useful purpose when properly designed and used.

The author finds the greatest amount of evapo-transpiration to date occurs from grass. In our own work we also find more evapo-transpiration from grass than from maise with which it was compared.

M. W. C. Lowdermilk said : Having had experience in establishing a number of batteries of lysimeters I should like to ask the author of this excellent paper if he had carried out a callibration period for the lysimeters. By this I should like to ask if all four lysimeters were treated alike for a number of years so as to establish the natural variability of the lysimeters as the basis for experimental comparison of different types of vegetative covers that are grown in the second stage of the experiment ?

I wish to commend the experiment as outlined and described by Dr. Deij.

Response of Dr. Deij : There has not been a callibration period for some years with bare lysimeters. As the lysimeters III and IV were seeded with leaf trees and needle leaved trees respectively they were in the beginning practically as bare as lysimeter I. The ratios E_{III}/E_{I} and E_{IV}/E_{I} for the first year (that means a half year after the beginning of the observation and a year after the seeding) are but slightly different from 1.

Dr. Slettenmark : — I would like to point out, that the value measured by lysimeter nº 2 agrees very well with the basin evaporation in southern Sweden, about 360 mm a year. The lower values in the other lysimeters as also the slow growth of the vegetation in the other lysimeters as compared with the surroundings may be, at least partly, explained by the fact that the soil has been plowed and, therefore, is more permeable than the natural condition.

Respons of Dr. Deij : Indeed we had some indications that in the first half year of the observations lysimeter IV, which was built last, showed some peculiar behavior with respect to the running off of the drainwater. Afterwards this disappeared totally.

Ir. B. A. van Nes indique que les lysimètres ont été construits avant la guerre, lorsqu'il y avait chomage. Le coût réel des lysimètres est de 100.000 florins.

Respons of Dr. Deij : The lysimeters were built before the war with the help of unskilled labourers which were in that time on the dole. The cost amounted about f. 100.000. They were partly born by other Water Supply Works.

Dr. van Nes présente ensuite le rapport préparé par the Netherlands State Institute for Water Supply.

M. J. Th, Thijsse : — Wishes to stress the practical importance of these investigations. The service of the Zuiderzee-reclamation has to know the quantity of water supplying the lake in midsummer, when the net rainfall is zero.

Because of the time lag between precipitation and flow into the lake of 90° the replenishment of the lake at that time is equal to the mean value of the precipitation.

M. Frolow rappelle que l'étude mathématique de l'écoulement qui avait déjà été bien avancée par Pavlowsky et par Hamel, aboutit à une solution complète dans les travaux de V. V. Vedernikov U.R.S.S. Un cas particulier a été reproduit ensuite dans un travail suédois, où la biographie est incomplète. Ces études permettent en particulier d'utiliser les modèles électriques.

Le Dr. van Nes répond que celles-ci n'ont pas été envisagées en Hollande.

M. Slettenmark : A similar investigation, regarding the flow of ground water to drain pipes in fine sandy, clay-free soil, has been made by Professor I. Gustafsson, University of Agriculture, Uppsala, Sweden. The laboratory experiments agree very well with the theoretical results.

Le Dr. Sayre présente ensuite les deux rapports des Indes qui ne donnent pas lieu à une discussion.

Le rapport du Dr. Gunnar Ekstrom est lu par M. Slettenmark.

Le Dr. J. E. Church prie M. Slettenmark d'obtenir de M. Ekstrom les réponses aux questions suivantes :

Why does the clay in the second stratum of the crust layer fail to swell from water in its fissures ? Is this due to age and hardness ? For drainage is used to remove the water from its base that has percolated thru the first stratum. Does the closing of the fissures in the surface stratum prevent the descent of sufficient water to fill the pores of the clay bed of the second ?

M. Le Strat fait observer que les sources sont plus facilement contaminables en été qu'en hiver par les infiltrations de surface survenant après une pluie importante. Les fissures du sol sont en effet béantes en été, alors qu'en hiver le sol gorgé d'eau est pratiquement imperméable.

G. W. Musgrave then presented his report on infiltration which contained the most outstanding developments in the subject of infiltration in the U. S. A. since the 1939 Assembly. A list of 31 of the most important papers is included.

M. E. Berkaloff lit le rapport établi en collaboration avec M. J. Tixeront.

M. J. Bertin ayant demandé comment tient-on compte du pourcentage de vides dans les massifs calcaires, M. Berkaloff répond que la méthode de Maillet employée envisage l'effet global d'épuisement des réserves souterraines et ne cherche pas à caractériser les terrains sauf par deux paramètres : coefficient de tarissement et volume initial.

A la question de M. Vibert y a-t-il en la circonstance et dans les cas étudiés égalité entre le volume d'eau écoulé du ou des massifs et le volume capté. M. Berkaloff répond que des écarts observés sont attribuables à l'évaporation. Ainsi pour 1948 l'écart serait de l'ordre de 30 %.

La séance est ensuite levée.

COMMISSION DES EAUX SOUTERRAINES

Réunion commune avec l'Association Internationale de Volcanologie
Tenue le 23 août 1948 à 14 h 30, sous la Présidence de M. Sayre

Le rapport de M. S. Guigue et G. Betier donne lieu à M. B. C. Escher de demander aux géologues, qui connaissent l'Algérie, si le souffre des eaux minéralisées ne pourrait être dérivé de la série gypsifère des terrains mésozoïques.

M. L. Glangeaud répond : Le volcanisme algérien s'est arrêté généralement au Miocène supérieur ou au Pliocène inférieur. Il paraît trop ancien pour se manifester même par des venues hydrothermales, sauf dans de rares exceptions.

La plupart des eaux minérales et thermales d'Algérie semblent provenir de nappes artésiennes plus au moins profondes retenues dans le fond des synclinaux ou synclinoriums. Le niveau le plus probable est soit le Lias calcaire, soit le Trias. Les flexures et fractures récentes d'âge pliocène supérieur et quaternaire jouent un rôle important dans la montée des eaux artésiennes profondes. Ce sont de véritables « sondages » naturels descendant jusqu'au niveau aquifère profond. En facilitant une montée rapide des eaux chaudes ces factures permettent le maintien d'une température suffisante. Il y aurait lieu toutefois de faire le bilan thermique de ces sources pour savoir s'il ne serait pas nécessaire, même avec ces fractures, de faire intervenir un flux de chaleur supplémentaire dû à des degrés géothermiques élevés.

Les rapports de M. H. Schoeller lus par M. Frolow, ne donnent pas lieu à une discussion, de même que le rapport de MM. Donald, E. White and W. W. Brannock, lu par M. Sayre.

Les rapports hongrois de M. F. Papp (lu par M. Frolow), Szalai, J. Sumeghy et J. Bogardi (lus par M. E. Mosonyi) montrent le beau développement acquis par l'école hongroise et la Commission, sur la proposition de M. Frolow, décide d'adresser ses félicitations aux hydrologues hongrois.

Les rapports de M. G. Dal Piaz, J. Healy et du Dr. W. E. P. Mc Lintock sont lus ensuite par leur titre.

M. Leslie E. Kent présente un très intéressant et très documenté rapport sur les eaux thermales de l'Union Sud Africaine.

M. Sayre approuve les remarques de ce rapport au sujet des termes.

Le rapport sur les eaux souterraines du Queensland, lu par M. L. E. Kent ne provoque pas de discussion.

Au sujet du rapport de M. Victor Jansa : Swedish Artificial Ground Water Supplies, M. A. N. Sayre remarque : The general use of artificial recharge for municipal supplies in Sweden is of extreme interest to the U.S.A. I would be pleased to receive a copy of any report on this subject which may be prepared.

La reponse est négative à la question de M. B.A. van Nes : are there no difficulties with clogging up of the bottom of the infiltration ditch ?

COMMISSION DES EAUX SOUTERRAINES

Séance du 26 août 1948 à 14 h 30, sous la présidence de M. A. N. Sayre

En ouvrant la séance M. le Président donne la parole à M. A. M. Vibert, dont le rapport n'est pas suivi d'une discussion, de même que celui de M. L. Tison qui précise seulement, sur la demande de M. A. N. Sayre que la valeur de S est de 4.5×10^{-3}.

Les rapports anglais sont présentés par M. C. S. Durst. Celui de F. Guyton l'est pas M. A. N. Sayre ; et celui de M. Castany, par M. J. Tixeront. Tous ces rapports ne sont pas suivis de discussion.

M. Mosonyi ayant présenté le rapport de M. A. Vendl, M. Le Strat demande quelle est la teneur limite en SO_3 compatible avec les besoins de la construction et obtient la réponse qu'elle est de 200 à 300 mg/l.

M. A. N. Sayre, président, lit ensuite les titres des rapports de MM. J. Healy, C. St. J. Mulholland, C. Cloe, R. Sarmiento Soto, S. Juan Hugo Ceasar, M. Gortoni, E. Nemeth, Hynie. M. Sayre donne lecture du rapport de C. C. Taylor sur les eaux souterraines du Chili.

M. A. Rittman expose le rapport de M. Wegenstein, après quoi M. L. Kent lit le rapport de M. Jansa. Au sujet de ce dernier M. V. Frolow donne un bref rappel des travaux français utilisant le procédé Schlumberger et M. A. N. Sayre indique le développement des sondages électriques en U.S.A., où beaucoup de succès ont été enregistrés.

Les rapports présentés étant épuisés, M. A. N. Sayre, président, propose à la Commission de se prononcer sur les questions pour l'Assemblée Générale de 1951. Après discussion des propositions faites par MM. A. N. Sayre, L. E. Kent, E. Mosonyi et V. Frolow, un vote a eu lieu et la Commission a adopté les trois questions suivantes :

1. — Run-off and ground water table, with special reference to forecasting.

2. — Variation du niveau et de la pression des eaux artésiennes (en collaboration avec les Associations de Géodésie, Météorologie, Magnétisme et Electricité terrestre et Océanographie).

3. — Source de chaleur et composition minérale de l'eau (en collaboration avec l'Association de Volcanologie).

Les travaux de la Commission ont été terminés, l'ordre du jour ayant été complètement épuisé.

Rapports relatifs à la première question : INFILTRATION

1) THE LYSIMETER STATION AT CASTRICUM (HOLLAND)

by

Dr. L. J. L. DEY
(Director of the Climatological Section of the Royal Dutch Meteorological Institut, De Bilt)

In the research of the evaporation of soil water the lysimeter has always been an important expedient, notwithstanding the fact that it has often violently been criticized. In the Yearbook 1866 of the Royal Dutch Meteorological Institute the then Director in Chief Prof. Buys Ballot already mentioned the use of small lysimeters of different depths up to 1, 2 and 3 dm, filled with soil and covered with grass. That in the Netherlands too the research of the evaporation of soil water was considered of high importance may be obvious from the fact that a prize was offered by the Utrecht Provincial Society of Arts and Sciences for the best answer on the question to produce : « Une série d'observations sur les quantités d'eau qui s'évaporent de divers terrains et de diverses plantes, dans les circonstances différentes ». The prize was awarded to the Englishman S. H. Miller in 1878, for his answer entitled « Prize Essay on Evaporation ». In this treatise experiments and results were described obtained with small cylindrical lysimeters of 20.3 cm diameter. They were filled with different soils (peat-moor, sand, clay, mould) and grown over with different crops (grass, clover and a few plants). Buys Ballot strongly recommended the use of lysimeters on a large scale « in order that one should obtain at length valid numbers at which machine power and seepage for reclaimed polders may be chosen and fixed ». Meanwhile the use of lysimeters remained restricted and their results were exposed to acute judgments. In 1897 the engineer Elink Sterk stated : « ... that the ordinary evaporimeters certainly don't indicate the evaporation from the free water surface and that the present lysimeter indicates highly inaccuralety the evaporation from the soil. Everyone who whishes to from a slightly correct notion of what happens in nature, should begin to dissociate entirely from the figures deduced from these instruments and forget them. Even to the question if the evaporation in one period be greater than in another, they answer always wrong. ... May they disappear out of memory as soon as possible ». In spite of this dark opinion lysimeters remained in vogue. Particularly agriculturists always attached much value to them as appears from the careful investigations which have been carried out by the State Agricultural Experiment Station and Soil Science Institute at Groningen since 1918. A detailed study was published by Ir. Maschhaupt in 1938. In addition to the Groningen experiments it contained a summary and a critical discussion of the principal foreign literature about the subject and on the whole it formed a pleading for « the good right of lysimeter research in answering questions related to the state of soil water ».

One of the most important objections against lysimeters is the disturbance of the natural structure of the soil when it is brought into the lysimeter. Especially this will be the case with stratified soils. If one has to do with a non-stratified soil then it will depend, as Maschhaupt states, upon the kind of soil if the disrupted natural structure will be of great influence on the movement of soil water. For sandy dunes and other light soils no objections are to be awaited in this respect.

Another objection that often arises against lysimeters is the absence of capillary connection between the soil in the lysimeter and the subsoil. This objection may be met without difficulty by maintaining an artificial water level in the lysimeter. A great advantage of lysimetres is the accurate measurement of the drainwater amounts.

Besides agriculturists there are the water supply engineers who had always a great interest in lysimeter experiments. It is for a water company a very important thing to have an exact knowledge of the evaporation in order to make a correct estimation of the amounts of water that are stored in different periods. Several water supply works, like those of the municipalities of Amsterdam and Den Helder, use them.

Already a long time the Provincial Water Works of North-Holland (P.W.N.) needed a lysimeter-installation. The P.W.N., in providing the major part of North Holland with drinking water, disposes of vast areas in the dunes between Bergen and Wijk aan Zee. Besides the P.W.N. has taken the laudable task to benefit the beautiful scenery in this region as much as possible by a strong enlargement of the scattered afforestation. Now the question arises how far,this afforestation will be profitable or not for the winning of potable water. If one will loose this problem with the aid of lysimeters than it will be clear that they must have such dimensions that a wood may grow up completely in it.

In 1936 the then Director of the P.W.N. Ir. van Oldenborgh took the initiative to establish a lysimeterstation near the principal pumping station at Castricum. In concert with the professors Jager Gerlings and Te Wechel of the Agricultural Academy at Wageningen one decided to build four lysimeters of armed concrete each with a surface of $25 \times 25 = 625$ m^2 and a depth of 2.5 m. Indeed this might be called a grand project and it inspires to great satisfaction, that in spite of many difficulties, the work was ready in 1941 and the regular observations began in the summer of that year.

The lysimeter domain that occupies an area of abour 90×270 m^2 is surrounded by a slightly hilly landscape of dunes. The domain itself is completly flattened. The four lysimeters from west to east numbereed with 1, II, III and IV are at distances sunk in the ground. Except number I, each has got its own planting that is continued in the nearest surrounding field. Upon lysimeter II one finds a normal lowly growing dune vegetation, consisting of grass, bent, thorn bushes, creeping willows and blackberry bushes. Upon lysimeter III oaks, alder-trees and birches were seeded and upon lysimeter IV pines planted with the aim that in the course of time they will grow up to a real wood.

In this way the influence of the gradual development of the soil planting on the usable precipitation may be followed just from the beginning. The lysimeters are filled with sand, partly from the digged ground, partly from the nearest surroundings. Old surface layers were carefully removed, only pure sand was used. The filling happenend as uniformly as possible by depositing in water. In order to maintain a constant ground water table a system of gutters is achieved at the bottom, the walls being 25 cm high. They are covered with shields with free small slits through which the water may flow off. A bedding of gravel prevents that sand is carried away. In this manner a horizontal groundwater table is guaranteed at a depth of 2.25 m below the surface in accordance with the mean water table in the surrounding field. The drainwater is collected in a small tank and from there it flows by a pipe-line into a large measuring tank of 5.5 m depth. The amount of drainwater which as flown off in one day is measured by means of a floating gauge. From time to time the measuring reservoir is emptied by an electrical pumping installation.

It may be clear that the lysimeter research only will be fully developped when the meteorological circumstances are considered too. With the co-operation of the Royal Dutch Meteorological Institute the lysimeterstation has been equipped with a comprehensive collection of meterological instruments. Firstly it is of great importance that the precipitation is accurately measured. Therefore several rain gauges are mounted in the field. The normal type consists of a cylindric reservoir with a conic bottom. Upon the reservoir a funnel is fitted with an opening of 400 cm^2. The rain gauge hangs in a fork with the rim at 40 cm above the ground. The amount of precipitation is measured in a calibrated glass. The most reliable results are obtained by means of a so-called ground rain-gauge, the rim of it being at the same level with the ground. To avoid splashing drops the funnel is surrounded by a brush mat which is hold in a wider grid. Finally a selfrecording raingauge of the tipping buckets type is mounted. Evaporation pans are placed near the lysimeters I, III and IV, in order to trace the influence of the growing wood upon the evaporation.

To the further instrumental equipment belong two thermometerscreens. The one is placed in the afforested domain, the other in the barren part. They contain a dry bulb and a ventilated wet bulb thermometer. a maximum and a minimum thermometer and a thermograph or hygrothermograph. At three places in the field soil temperatures at 1, 10, 25, 50 and 100 cm are measured.

The wind velocity is recorded by a Robinson anemometer at 9 m above the ground and the wind direction is indicated by a vane.

Finally the sunshine duration is recorded by a Campbell-Stokes instrument. The climatological observations are performed at 8, 14 and 19 o'clock (local time) by an observator and an assistent. They have a small brick house at their disposition where the observations are worked out and where tools and auxiliary instruments are put away.

It may be called a happy circumstance that, in spite of the german occupation, the observations could be continued almost unimpeded and that the lysimeter installations and the instrumental equipement have remained untouched. The only dis, agreeableness was the fact that during the winter half year the observation at 19 o'clock cancelled on account of obscuration measures.

Up to now it has not been possible to execute the plan to determine the evaporation in relation to the meteorological circumstances on account of the lack of the needed instruments and the difficulties experienced with the manufacturing of them during and after the war.

Merely the results of the annual evaporated amounts of water as measured by the four lysimeters may be mentioned.

We may set up for each of the lysimeters the water balance ,valid for an arbitrary chosen period :

$$P = D + E + S_2 - S_1$$

Here P means the amount of precipitation, D the amount of drainwater, E the amount of evaporated water, S_1 the amount of water stored in the soil at the beginning of the chosen period and S_2 the corresponding value at the end. Generally S_2 will differ from S_1. The quantities P and D are daily measured but S_1 and S_2 are unknow. For a sufficient long period, when the soil moisture content at the beginning is but little different from that at the and, like it is case in the late fall, the difference $S_2 - S_1$ may be neglected with respect to P — D and then the evaporation may be determined with a sufficient accuracy.

In the subjoined table the annual amounts are summarised over periods 1 December — 30 November for lysimeter I (bare), lysimeter II (partly covered with a natural dune vegetation), lysimeter III (overgrown with a leaf-wood) and lysimeter IV (overgrown with a needle leaved wood). The amounts of P, D and E are all expressed in mm water-height. Further the evaporation values, expressed in percents of the annual precipitation are determined and finally the ratio's of the evaporation from the overgrown lysimeters to the bare one.

1942

		I	II	III	IV
P	853.1				
D		708.5	613.8	696.2	686.1
E		144.6	239.3	156.9	167.0
$\frac{E}{P} \times 100$		17.0	28.0	18.4	19.6
$\frac{E}{E_I}$		1.00	1.65	1.08	1.15

1943

		I	II	III	IV
P	707.2				
D		500.3	351.7	440.0	458.2
E		206.9	355.5	267.2	249.0
$\frac{E}{P} \times 100$		29.3	50.3	37.8	35.2
$\frac{E}{E_I}$		1.00	1.72	1.29	1.20

1944

		I	II	III	IV
P	889				
D		690.4	546.6	599.8	617.6
E		198.7	342.5	289.3	271.5
$\frac{E}{P} \times 100$		22.3	38.5	32.5	30.5
$\frac{E}{E_I}$		1.00	1.72	1.46	1.37

1945

		I	II	III	IV
P	825.6				
D		654.8	503.7	555.3	552.2
E		170.8	321.9	270.3	273.4
$\frac{P}{E} \times 100$		20.7	39.0	32.7	33.1
$\frac{E}{E_I}$		1.00	1.88	1.58	1.60

1946

		I	II	III	IV
P	872.8				
D		651.2	512.1	562.7	534.6
E		221.6	360.7	310.1	338.2
$\frac{P}{E} \times 100$		25.4	41.3	35.5	38.7
$\frac{E}{E_I}$		1.00	1.58	1.40	1.53

1947

		I	II	III	IV
P	652.8				
D		411.7	275.5	308.3	262.3
E		241.1	373.3	344.5	390.5
$\frac{P}{E} \times 100$		36.9	57.1	52.7	59.9
$\frac{E}{E_I}$		1.00	1.55	1.43	1.62

The comparison of the results leads to the following qualitative conclusions. In 1942 the evaporation values for the lysimeters III and IV are only slightly different from the value for I, but lysimeter II shows a rather strong evaporation. In the next years the evaporation in the overgrown lysimeters generally increases. As compared with the bare lysimeter the increase of the evaporation is strongest for IV, a little less for III and still less for II. This general course may be declared from the nature and the behaviour of the vegetation.

The total evaporation may be divided into three parts : 1° direct evaporation from the soil by means of its pores, 2° direct evaporation of the precipitation caught by the vegetation, 3° evaporation of soil water via the plants by means of their organs. This last form of evaporation is commonly called transpiration. As the plants are growing up the transpiration will increase. This behaviour is clearly reflected by the values of the total evaporation. The direct evaporation of the precipitation caught by the growing vegetation will increase also. But the evaporation of the soil will diminuate as a consequence of the decreasing insolation and the weakening of the wind velocity near the surface of the soil. As long as the leaf trees and the needle leaved trees are of small demensions the total evaporation from the lysimeters III and IV will but slightly deviate from that of I and the differences must be chiefly imputed to the transpiration. The high value of the evaporation from lysimeter II may be explained by the direct evaporation of the precipitation caught by the grass cover.

As already remarked a more detailed researh for the evaporation in relation to the meteorological circumstances was not yet possible. The principle purpose now will be the determination of the soil moisture content. Several methods are at our disposal. A well known method, particularly used by agriculturists, consists in sampling the soil by means of a ground drill. But a manyfold application of this method seems undesirable as the structure of the soil will be disturbed at length. This danger is avoided by using tensiometers. They consist of a small tube of porous clay, filled with water and connected with a mercury manometer. The tensiometer, placed at a fixed depth in the soil marks a pressure depending on the moisture content of the soil and on its structure. The slowness of the instrument is a disavantage. Another method which will be used is based on the relationship between the conduction of heat and the soil moisture content. By means of an electrical heat source at a fixed depth in the soil a small amount of heat is supplied to the surroundings and the variation of the temperature is measured. With the aid of theoretical considerations the heat conductivity is calculated. Separate experiments will establish the relation between heat conductivity and soil moisture content.

2) REPORT ON « EVAPORATION FROM BARE SOIL IN RELATION TO THE DEPTH OF THE WATER TABLE OR THE ZONE OF SATURATION »

by

Dr. L. A. Ramdas,
Director of Agricultural Meteorology, Meteorological Office, Poona.

1. Introduction

The Agricultural Meteorology Section at Poona has been engaged, amongst other activities, in investigating the movement of moisture through the soil, evaporation from free water and soil surfaces, effects of certain salts on the permeability of soils to water, etc. (1 to 11). These investigations, which are in progress at the Central Agricultural Observatory at Poona, have been briefly reviewed in a report by the present writer presented to the Research Committee of the Central Board of Irrigation in India in July 1947 (12).

The experiments conducted at the *Central Agricultural Meteorological Observatory at Poona* show that the evaporation from the soil surface is comparable in magnitude to that from a free water surface so long as the soil surface is saturated with water. When the soil dries up, the evaporation will depend also on the upward or capillary ascent from the wetter soil below. When this supply also diminishes, the soil surface begins to exchange water vapour with the air layers above in the manner described in a series of papers from Poona (13 to 19).

In the present note we shall discuss the interesting results obtained with a series of soil evaporimeters containing soil layers of various depths with their bottom resting on a water table. They suggest a method of estimating evaporation losses from the bare soil from a knowledge of the depth of the zone of saturation below the surface.

2. The Soil evaporimeter

The instrument (see Fig. 1) consists of a metallic cylinder ABCD 5 inches in diameter for holding the soil under investigation. The soil is packed uniformly up to the top of ABCD and rests on the perforated bottom at CD. The cylinder ABCD dips into the reservoir EFGH which contains water. Water can be poured into the reservoir through the inlet at I which is provided with a screw tap. The level NO of water in the reservoir may be observe in a glass side tube J at M. The side tube is provided with a cap to prevent evaporation. At KL where the cylinder ABCD enters the reservoir an adhesive tape is attached all round to prevent any loss by evaporation.

For the study of the effect of varying the distance between the water table and the top of the soil column on the loss of water by evaporation, a series of instruments with soil columns of 6″, 1ft, 1.½ ft, 2 ft and 3 ft (corresponding to AN in Fig. 1) are exposed in the open. Special care is taken to see that the packing of the soil into the vertical cylinders of the evaporimeters is uniform for all instruments the weight per inch of the soil column being adjusted to be the same when the packing of the soil is done in stages. Three soils could be experimented with at a time as three sets of soil thermometers of the sizes referred to above were constructed. To make the observations strictly comparable, uniformity of exposure is secured by supporting all but the largest size (3 ft) on suitable stools so that the tops of the soil columns are all at the same level (3 ft 3″) above ground.

3. Measurements

The evaporimeters are weighed daily at 0700 hours local time, replenished with water up to the reference mark on the glass side tubes and weighed again. The weighing is done correct to 1/8 ounce. The difference between the *weight after adding water on one day* and the weight *before* adding water in the next day is the actual loss of water by evaporation from the soil surface. To convert the evaporation in pounds to inches, the loss of weight in pounds is multiplied by the appropriate conversion factor (1.41).

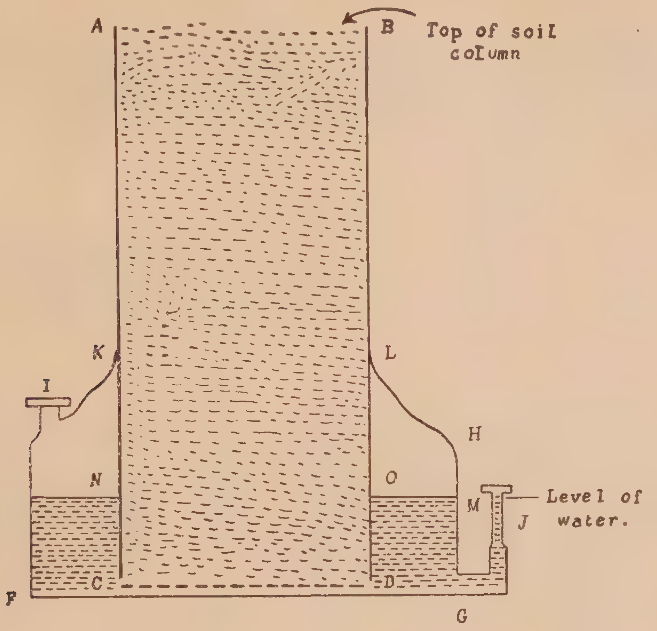

Fig. 1 — Sketch of soil evaporimeter

Some typical examples may be quoted here. In one of the experiments the first set of evaporimeters was packed uniformly with the black cotton soil of Poona, the second set with normal soil of the Punjab and the third set with Punjab « Bari » (or Alkali) soil. The time taken by water to rise to the top surface of the soil columns as indicated by the appearance of wetness and also as verified from the daily evaporation losses, was noted. Table 1 gives the time taken by water to rise up to different heights in the three soils referred to above.

Table 1. — *Time of ascent of water.*

Type of soil	Time taken by water to ascend to a height of				
	6″	1 ft.	1½ ft.	2 ft.	3 ft.
Poona Soil	1 day	2 days	6 days	15 days	Surface not perfectly wet even after six months.
Punjab Soil (normal)	1½ days	3 days	5 days	10 days	21 days
Punjab « Bari » Soil	3 days	(Surface not wet even after six months)			

The evaporation losses became more or less steady once the initial period of ascent viz. December 1938 was over. The mean daily values of evaporation in inches, during January, February, March, April, May and June 1939 are given in Table 2.

From Table 2 it will be clear that

1) in all soils the evaporation decreases as the water table recedes below the evaporating surface,

2) the evaporation from the « bari » soil is very much smaller than that from either the Poona soil or the Punjab « normal » soil,

3) when the water table recedes below 1 ½ ft from the surface, the evaporation from the more porous alluvial « normal » soil of the Punjab is more than from the black cotton soil of Poona which has a larger clay fraction and

4) as is to be expected, evaporation increases with the advance of summer and then decreases with the advent of monsoon conditions.

Table 2. — *Mean daily evaporation.*

Month (1939)	Type of soil	Mean daily evaporation in inches when the distance between the evaporating top surface of the soil column and the water table is				
		6″	1ft.	1½ ft.	2 ft.	3 ft.
January	Poona soil	0.30	0.26	0.17	0.11	0.01
	Punjab « normal » soil	0.29	0.23	0.21	0.18	0.09
	Punjab « bari » soil	0.05	0.02	0.02	0.01	0.01
February	Pooàna Soil	0.38	0.34	0.22	0.15	0.02
	Punjab « normal » soil	0.37	0.26	0.22	0.18	0.09
	Punjab « bari » soil	0.06	0.02	0.02	0.02	0.01
March	Poona soil	0.48	0.41	0.23	0.16	0.02
	Punjab « normal » soil	0.43	0.30	0.21	0.17	0.08
	Punjab « bari » soil	0.05	0.03	0.03	0.02	0.01
April	Poona soil	0.48	0.40	0.23	0.16	0.01
	Punjab « normal » soil	0.51	0.31	0.23	0.18	0.09
	Punjab « bari » soil	0.05	0.03	0.03	0.02	0.01
May	Poona soil	0.49	0.39	0.21	0.14	0.03
	Punjab « normal » soil	0.51	0.32	0.24	0.18	0.08
	Punjab « bari » soil	0.04	0.03	0.03	0.02	0.01
June	Poona soil	0.22	0.09	0.08	0.07	0.02
	Punjab « normal » soil	0.22	0.16	0.13	0.12	0.07
	Punjab « bari » soil	0.03	0.03	0.02	0.02	0.01

Experiments similar to the above have been made with different soils and in each experiment one set of evaporimeters was filled with the Poona soil which was always used as a standard. Evaporation values are now available for the red soil of Bangalore, the sea sand from Trivandrum, the black soil from Coimbatore and the clayey soil of Jubbulpore. The experiments are being continued with other soils during the clear season at Poona. To try out a larger number of soils rapidly, more sets of soil evaporimeters are being used set up.

4. Discussion : evaluation of the « α » Coefficient

We have seen that the evaporation from a soil surface decreases rapidly with distance between the water table and the evaporating surface. From the mean values of evaporation similar to those shown in Table 2, the relation between the evaporation E_z and the distance Z (in cm) between the water table and the evaporating surface is found to be of the form

$$\frac{E_z}{E_o} = 10^{-\alpha z}$$

where E_o is a constant which is equal to the evaporation when Z is equal to zero and α is another constant characteristic of the soil. The ratio of evaporation when Z is equal to say 50 cm to that when Z is say 20 cm is then given by

$$R = \frac{E_{50}}{E_{20}} = 10^{-\alpha(Z_2-Z_1)} = 10^{-30\alpha}$$

The values of α can be computed readily by plotting log R (keeping the denominator equal to $E_{(15.24\ cm.)}$, the numerator taking the other known values of E_z) against Z. The curves are found to be linear.

The values of α for the various soils experimented with so far are given below:

Table 3

Name of soil	α (when Z is in cms)
Poona black cotton soil	.014
Red soil of Bangalore	.015
Alluvial soil of the Punjab	.0083
Jubbulpore soil	.010
Coimbatore soil	.0096

The importance of determining α experimentally for as many typical soils as possible will be realised when it is observed that when α is known and estimates of E_o (with instruments flush with ground level) are available, it will be possible to make reasonable estimates of the actual evaporation losses from large catchments in different seasons. The only additional information required for this purpose will be the depth Z of the water table or level of saturation in the area under consideration. Estimates of Z and its seasonal variation can be made at selected localities.

Further work on this subject is actively in progress at the Central Agricultural Meteorological Observatory.

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3) INFILTRATION IN RELATION TO PERMEABILITY IN SWEDISH SOILS

by

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The total precipitation in Sweden averages about 500 to 600 mm annually but in a few places it exceeds 1,500 mm, while certain places have an annual precipitation of only about 400 mm. On the West Coast the precipitation is comparatively high — about 800 mm annually.

Generally speaking ,more than half of the precipitation taking place in Sweden evaporates. The rest finds its way, in one way or another, to the sea. Part of this flows off superficially on the surface of the ground as runnels, rivulets and the like, while the rest sinks down into the ground and gradually, as ground water, runs through premeable earth or rock layers into some lake or watercourse, or comes to the surface on the form of springs, or flows directly into the sea.

The volume of water evaporated from a given land area has been calculated as the difference between the annual precipitation within the area and the amount of water drained off from the area during the year. In the evaporation, is included the loss of water from land and lake surfaces and the water escaping through the vegetation. On the average, the evaporation from a large land area in Central and Southern Sweden amounts to about 360 mm (Wallén, 1934). The evaporation is on the whole, constant from one year to another — in other words, it changes very little with the amount of precipitation. In a year of high precipitation this, consequently only affects the drainage, which is then increased.

The drainage has been calculated by measuring, by one method or another, the amount of water flowing through watercourses during the year. The entire drainage does not, however, take place directly through watercourses. Quite large amounts of water are accumulated for longer or shorter periods in lakes or in earth and rock layers, as ground-water reservoirs. A large part of the water drained off has thus come part of its way in the form of ground-water streams.

As the average annual precipitation is 550 mm and the evaporation 360 mm, the remaining part of the precipitation would be about 200 mm. About one half of this can reasonably be assumed to percolate into the ground and become ground water (Ekström, 1941). The remainder runs off superficially on the surface of the ground.

When selecting a place for a ground-water supply, it should be seen to, that it has the largest precipitation area possible. The larger the collecting area of the precipitation, the more plentiful the ground-water supply to be expected. When the surface-water divide has been stepped off, and traced on a map, the size of the precipitation area can be calculated. On the assumption of the precipitation sunk into the ground being 100 mm per year, a ground-water supply of about 115 litres per hactare (10.000 m^2) and hour can be expected. By multiplying the precipitation area, expressed in hectares, by 115, it should then be possible to calculate the approximate water supply in litres per hour which should be available from a suitable placed well, within the precipitation area.

This method of calculation is, however, subject to several sources of error and can only be applied to large precipitation areas. The greatest sources of error are found in the fact that the amount of infiltrated precipitation varies greatly, owing to differences in the porosity or permeability of the soil, and that the ground-water divide and the surface-drainage divide, do not always coincide. The declivity of the ground must also be considered, as topography in many cases plays a very important role.

The sinking, or infiltration, of the precipitation into the ground depends to a great extent on the permeability of the surface layers of soil and on the prevalence of fissures in the bedrock surface. On gravelly and sandy soils the surface drainage is practically none, and the infiltration consequently great. These soils are known as dry soils which are not drained. On clay soil the water flows off superficially or through drain pipes, unless infiltration takes place in adjoining, sandy areas In a clay soil the infiltration is thus insignificant.

On the occurrence and utilization of ground water for consumption purposes, a report was presented at the conference of the section in Washington in 1939 by Doctor Simon Johansson. A more detailed report was later presented by the present writer (Ekström, 1941).

The freezing of the ground during the winter months naturally has a considerable influence on the water infiltration. It reduces the permeability of the soil more or less, The formation of ice in the ground depends on the water content in the soil, the upper level of the ground water and the character of the soil (Beskow, 1935). In densely wooded aeras the frost stays longer in the ground. When the frost disappears, the permeability of the soil is increased as a result of the development of frost fissures.

Freezing may influence the stability of the structure of the humus-containing surface soil in different ways. High content water, repeated freezing, and freezing at low temperatures, tend to produce a finer -grained structure. A moderate water content and slow freezing near the freezing point result in a coarser and more stable structure (Torstensson and Eriksson, 1942).

Rain causes a temporary settling and compression of the surface soil. This settling is greatest in the lower part of the surface soil and decreases gradually towards the surface, and in the top layer of about five centimeters no compression occurs. The settling is caused by swelling of the clay aggregates, through which the cohesion is lessened, and the earth lying on top compresses the aggregates (Johansson, 1943).

Infiltration is affected by the mechanical composition of the soils. Soils with a low clay content, but with a high content of very fine sand (0.06-0.02 mm) or silt (0.02-0.002 mm), are affected by rain, and the surface of the soil becomes dense and almost impermeable. In such cases the run-off is pretty high.

On cultivated ground the degree of looseness in the surface soil is very important for the infiltration. The tilling of the soil here plays a prominent part. In the surface soil of a fallow, the degree of looseness can thus be about 37 per cent, and in the third year of hay field about 19 per cent (Johansson, 1933). A fallow thus retains a greater part of the precipitation. On old hay fields much water runs off superficially on the surface of the ground. As a protection against the — particularly in Eastern Sweden — usual spring drought, the soil should in spring be worked as little as possible in order better to retain the moisture. Ploughing should be done in the autumn. Deep tilling in the spring is a mistake from the moisture point of view.

The permeability of very well-assorted, coarser sediments is directly proportional to the square of the grain size (Beskow, 1935). In our ordinary soils, which are not so well-assorted, the permeability depends on the mechanical composition that is the degree of assortment. It is also influenced by the degree of packing and the grain-shape.

In soils with little or no clay content the ground water and the sinking free water ,move in the same manner through the pores throughout the whole layer. A laboratory test — by means of a permeameter, an oedometer, a compressometer or some similar instrument — gives sufficient accurate measure of the permeability of the soil. Tests of the effective size of grain and the uniformity coefficient according to Hazen, have been made in order ot ascertain the water-carrying capacity of sand and gravel layers (von Greyerz, 1925).

Gustafsson (1946) has studied the movement of groundwater towards drainpipes in finely-grained, sandy, clay-free soils. By means of experimental investigation and field tests, as well as by theoretical calculation, he has made clear the stream lines of the ground water around the drain pipes.

Among the clay-free and clay-poor soils, the most common moraine soils have a unique position, as they were originally deposited by the continental glacier and thereby became unassorted, the layers forthermore being hard and compact. Their permeability is therefore low. Down to a depth of about 0.5 meter, however, the top layers or the so-called surface moraine is fairly permeable as a result of the effects of frost and plant roots, and because this layer already during its formation acquired a coarser structure than the underlying ground moraine. In the latter the water chiefly moves through certain diffuse water ducts (ground-water arteries). These arteries mainly seem to follow the bedrock surface (Tamm, 1931). Gravelly and coarse -grained, sandy moraines are more permeable than those consisting mainly of fine sand.

In clay-free, coarse-grained soils the permeability is, as already stated, the same everywhere, and the soil is homogeneous from a hydrological point of view. In clay-soil, on the other hand, the pores between the grains (the textural pores) are so fine that any flow of free water is out of question. The permeability of a clay soil thus depends entirely on interstices (shrinkage fissures, worm holes and root canals) existing in the clay. The permeability is dependent on the size and durability of the fissures and on how interconnected the fissure system is. A fissure-free clay is practically impermeable to water.

The upper layers of a sedimentary clay, deposited in water, are more or less dried out. By digging down into it one finds, however, that the clay with increased depth becomes more and more soft and saturated with water. Downwards from a depth of, as a rule, two metres below the ground surface it is very soft and of a soapy consistency, so-called soap elay, usually bluish grey in colour and therefore colloquially referred to as « blue clay ». The part of the clay lying above the soap clay constitutes the so-called dry crust. From the point of view of permeability this can be divided into two layers, the upper and the lower dry-crust layer (Ekström and Flodkvist, 1926).

The upper dry-crust layer (that is the upper 60-80 centimeters of the dry crust) is in regard to its permeability dependent on the periodical drying,water-soaking and freezing of the ground. During summer this layer is dried out to a greater or lesser degree and interlaced with rather wide shrinkage fissures, by which water as well as air easily penetrate. During spring the permeability is also somewhat greater than, for instance, in the autumn, owing to the frost fissures. During a period of high precipitation, on the other hand, these fissures close, owing to the clay swelling and oozing down, and their duration is therefore wery short.

The lower dry-crust layer seems, on the other hand, to be unaffected by the annual meteorological changes. The clay here retains, on the whole, the same water content and the same consistency acquired by it during the drying-up in connection with the land elevation which took place several thousand years ago. It was thus during this period that the fissures in the lower dry-crust layer developed. A characteristic of these fissures is that they are permanent and that they have the form of distinct, vertical fissures which divide the clay layer into columns with a width of about one decimeter each.

The upper dry-crust layer thus has temporary fissures, whereas the fissures in the lower dry-crust layer should be stable and permanent. As against this view advanced by Ekström and Flodkvist, Johansson has, in regard to the fissures in the lower dry-crust layer, expressed a different opinion (1938). He thus holds, on the basis of measurements made by him at varying depths below the ground surface in clay soils with low permeability, that these fissures cannot be permanent. At a depth of, for instance, about 1.6 metres, the fissure volume in March, in water-saturated soil was 2.0 per cent, and in September, when the soil was driest, 9.4 per cent. As the water content of the clay did not differ greatly, it seemed that the difference in volume could not have been caused by shrinkage. Johansson therefore concluded that, owing to the solidity of the earth-mass, the development of fissures caused by shrinking in the upper layers continues further downwàrds than the loss of water allows for, as a sort of explosive effect, causing the layer to expand up wards and the fissure volume to incrasee. With increased moisture the process is reversed. The swelling pressure causes pressure on the lower layers, and the fissures are compressed in a strictly mechanical manner. The fissures would then be variable in width and, consequently, in their permeability. In the upper dry-crust layer the fissure volume is conditional, not only on shrinking and swelling, but also on the seeping down of material from the cultivated soil. This is greatest when the cultivated soil has a powdery structure, as when root crops are grown.

The fissures, as well as the root canals and worm holes, are depositories for the ground water in the lower dry-crust layer. If these formations did not exist, the clay would be completely impermeable to the sinking water. The fissures are, however, usually very narrow, and their width often does not exceed 0.5 mm; they are, furthermore, in some sections wholly or partly clogged by seeped-down clay substance. The ground-water reservoirs to be found in clay subsoils are consequently not very great.

By reason of the fissures being narrow and partly clogged up, the ground water is only slowly drawn to the drain pipes. Clay soil, therefore, as a rule has low permeability. Heavy precipitation and melting of snow causes the fissures to fill quickly with water, and as a result the ground water rises up into the surface soil, which beceomes strongly saturated.

Flodkvist (1931,1935) divides the ground water into temporary and permanent ground water. In cultivated soil the former appears in at surface soil which rests upon subsoil with low permeability, and is therefore called surface-soil groundwater. The fissures in the subsoil being filled with water, the surface-soil groundwater consequently has to find its way through the lower part of the surface soil until it reaches the re-dug earth above a drain pipe, through the cavity of which it reaches the drainage system. The re-dug earth above the drain pipe lets the water through considerably better — about one hundred times faster or more — than the earth between the drains.

On sloping ground, the run-off is of course much quicker than where the ground is level. Transversal draining is therefore decidedly more affective than longitudinal draining.

Ordinary clay soil is in most cases not very permeable to water ,and therefore water soaked and in great need of draining. In certain, usually small, low-lying areas of clay or gyttja, the fissures in the lower dry-crust layer can be comparatively large — from a few millimeters to 10 centimetres in width; the upper dry-crust layer is, furthermore, of a grainy consistency. Such areas are to be found here and there in the country, and it turns out that the larger fissures have developed as the result of a comparatively speedy drying out of a clay or gyttja subsoil in connection with the lowering of lakes, ditch-draining of low-lying, marshy land, reclaiming of land by dyking etc. Where the subsoil has such large fissures, draining is entirely superfluous, as the fissures constitute a more than adequate substitute for grain pipes. The open ditches can here be placed far from each other.

A characteristic of the self-draining subsoils is that they have a heigh percentage of clay, gyttja and dy (mud). These colloidal substances have the capacity of binding considerable quantities of water; they often hold as much as 75 per cent by weight of bound water. The self-draining are usually gyttja-containing clay, clayey gyttja and gyttja.

But also in other highly colloidal subsoils found in low-lying areas, draining gives rise to more or less wide fissures, as for instance in humus-free, glacial clay. From the typical, wide-fissured, self-draining subsoils, to the hardly permeable clay subsoils, all intermediaries are thus to be found (Ekström, 1947).

In the glacial, heavy clay found on the Lanna Experimental Farm there are fairly wide fissures, making it possible to lay the drain pipes rather far apart 24 meters or more (Perman, 1946). Investigations made by Johansson (1944) showed the fissure volume in the lower dry-crust layer, in the beginning of June and at depths of 1.1 and 1.6 meters, to be 8 and 4 percent respectively. The ground water level at that time lay 0.8 meter below the surface of the ground. The fissure formation seems to have developed as a result of as speedy lowering of the ground-water level in connection with the digging of deep drainage ditches.

In this connection Johansson worked out a new field method for determining permeability. Water was pumped out of a bore hole, and simultaneoulsy observations were made of the lowering of the water level in other bore holes within the sinking funnel resulting therefrom. This pointed to the conclusion that the permeability of the Lanna clay approximates that of fine sand of 0.2 mm grain size.

An approximate idea of the permeability can be had by ocular examination of the fissure development in a pit, or by an estimation of the inflow of water in a pit or a bore hole.

A run-off of ground water from the drainage systems occurs only after the melting of snow or after heavy rains, Exceptions are fields in which infiltration of ground water, sometimes in the from of springs, takes place from adjacent areas.

Flodkvist has (1931), in five different experimental fields in Central Sweden studied the permeability of different soils and the run-off of ground water from the drainage systems in connection with precipitation, the declivity of the ground, evaporation etc. A run- off of 1-3 litres per second per hectare is not uncommon, but then only for a few hours at a time. The highest run-off messured was 6.0 litres.

On the Russelbacka farm in Västergötland, with clay soil of low permeability, Johansson has (1938) measured the run-off from six different drainage systems, and also the height and variations of the ground-water level at three different distances from a drain pipe. At the same time observations were made of the precipitation and the depth to which the ground froze in winter.

When the moraine clays were formed, they were compressed by the continental ice-cap, which was some kilometers in thickness, and thus became exceedingly compact. No further shrinkage is therefore possible, and no drying fissures can develop in them. The fissures and cavities wich appear in them, are evidently the result of alternate drying and water -soaking of the surface layers, and of freezing, and they reach no great depth. The moraine clays are therefore rather compact and need comparatively heavy draining.

In moraine clay, fissures are, it is truc, quite frequently found at various depths below the ground surface. These fissures are not drying fissures or open fissures, but a kind of gliding surfaces developed by pressure when the clay was compressed under the ice. The movement of water in the moraine clays seems to take place chiefly through the more sandy and gravelly lenses which normally occur spread throughout out the moraine (Ekström, 1934).

In regards to peat soils, the rule is that peats which are only slightly mouldered and at the same time loose, let the water through rather easily, while on the other hand the permeability of more completely mouldered and compact peats are exceedingly low. In the slightly mouldered peat soils, the permeability seems to be five times greater in a horizontal than in a vertical direction (Lalmström, 1928).

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4) COMMISSION ON UNDERGROUND WATERS
INFILTRATION

by

G. W. MUSGRAVE, U.S.A.

Of the nine years intervening since the last meeting of the International Union of Geodesy and Geophysics, more then half of them were primarily devoted in America to war activities. Research in most fields, except those of military nature, was greatly curtailed. It often seemed, from a comparison of developments from one year to the next, that there was little evidence of progress. Such particularly was true of research in the subject of infiltration.

When we contract our knowledge at the begeinning with that at the end of the period, however, we see that actual progress has been made in our understanding of the fundamentals of the phenomena governing infiltration. Advances were also made in the practical application of the infiltration concept to various field problems.

These advances have come not so much from any single piece of research as they have from an overall comprehension of numerous separate studies conducted by many agencies and individuals. Some things that had been doubtfully or tentatively set forth or were vaguely understood have been clarified. Many separate pieces of information have been found to fit together nearly to form a larger segment of the solution to problems of how, when, why, where, and how much certain factors affect infiltration.

About the time of our last meeting in 1939, it was not uncommon to regard the rate of infiltration as a linear function of time (15). We had learned that rates differed for various soils and for different vegetal covers, but the reasons for these observed results were at best only dimly seen. We were beginning to recognize the possible utility of the infiltration concept in the prediction of surface runoff but, even at a later date, the supposed merits of « runoff coefficients » as a basis for estimating the amount of runoff still were being advocated (14).

During the past decade sufficient experimental evidence has been obtained to permit us to outline the major causal and modifying factors of infiltration. As concisely as possible, and perhaps with some over-simplification, the process as we now see it will be outlined.

In a broad way we are dealing with the interaction of a fluid and a permeable medium. In so far as the fluid and the permeable medium are constant and uniform in their characteristics, the process is relatively simple.

The prevailing size and numbers of pores govern the resistance to flow and the head of water governs the force. The viscosity of the fluid is recognized and Darcy's Law applies (21).

The problem of infiltration of water in soil becomes complex, hewever, because neither the fluid nor the permeable medium is constant and uniform. Sands are relatively more stable than most soils. Satisfactory working formulae governing the flow of water (and other fluids) through sands have been developed (6) (18) (26). These essentially are Darcy's Law in one or another of its possible forms. In contrast, the prevailing sizes and numbers of pores in soils are constantly hanging, and in most instances the applied water also changes in chemical and physical characteristics during the course of infiltration.

It is in the fine-textured soils that the greatest changes occur in structure. The coarse-textured sands and sandy loams are much more stable than such soils as our aggregated silt loams.

What are the means by which an aggregated structure is formed ? Under what circumstances is it destroyed ?

Aggregate Formation and Permeable Structure

Although we need to know about the process of aggregate formation in soils, we can now see at least its outlines. To form a soil aggregate there is needed a cementing agent or binding material, and a force to produce pressure. Some of the cementing agents are various salts precipitated out of solution, various colloids both organic and inorganic in source, possibly the strands of mycelium of fungi and also the mucus that may be derived from these or other sources.

An example may be cited from one of McCalla's studies (19). Using Peorian loess from a horizon practically devoid of organic matter, he applied a weak solution of leachate of decomposing plant residues. To another portion of loess he applied an equal quantity of pure water as a check treatment. Both were dried and small aggregates or lumps of each subjected to the impact of drops of water falling from a specified height. The relative stability of the aggregates or lumps was determined be the number of impacts before disintegration occurred and was found to be many-fold greater for the organic leachate than for the water. The method of study, in addition to clearly proving that the soluble organic matter is a means of building water-stable aggregates, offers opportunities of elaboration of concepts of causal factors.

Another method of forming water-stable aggregates has recently been shown by Hopp and Hopkins (13). Taking the lead offered by Darwin's studies of earthworms, it has been shown that (a) the earthworm casts are more stable in water than simple lumps of the same soil; (b) the numbers of water-stable aggregates in a soil are correlated with the number of earthworms present; (c) the size of the aggregates is related to the species and size of the earthworm; and (d) that of the various factors conducive to a large earthworm population, organic matter is one of the most important. The studies have also thrown light on other factors, affecting an earthworm population, such, for example, as temperature, the soil reaction, and the presence of nutriente. It appears that in the earthworm the complete mechanism for aggregate formation is present, pressure being provided as the soil passes through its body, while mucus and possibly other cimenting agents are supplied simultaneously.

Many other examples of aggregate formation could be described; the more outstanding being included among the literature cited (1) (10) (16) (30).

The numbers, sizes and relative stability of aggregates apparently varies with the kind of soil organisms that are present. The early findings of the Russian investigators, indicating that generally grasses are superior to legumes in their effects upon structure, apparently have been confirmed. Hubbell, as well as McCalla, has established that the species of microorganisms prevalent in the soil dominantly control the numbers and kinds of aggregates that are formed.

To generalize very broadly from the work to date, the essential features of soil structure in relation to infiltration may be indicated as follows :

1) Achieving a well aggregated soil structure is one of the most feasible ways of obtaining large pores and improved infiltration in finely-textured soils.

2) To be most useful in improving infiltration of water, the aggregates should have considerable stability in water.

3) Soil aggregates occur most commonly where there is present some clay, colloid, and organic matter.

4) The proportion of the soil that is aggregated, if effective upon rate of water, movement, must be so large that the volume of the unaggregated portion is less than the volume of voids between the aggregates — that is, the proportion of particles of large, medium, and small size must be such as to favor *porosity*, not *density*, of the mix.

Destruction of Aggregates and Deterioration of Permeability

The processes of structural formation in virgin soils are of the general character indicated, but conditions differ from those on cultivated areas largely in the surface protection that is provided by a canopy of forest or grass or other continuous vegetation above the soil. The favorable structure that is formed is thus largely preserved. In contrast, on cultivated or bare areas it is exposed to several deteriorating or destructive influences.

The impact of raindrops upon any exposed soil surface is one of the most common and destructive forces tending to break down soil aggregates and convert a permeable surface soil horizon to an impermeable one or one that is less permeable than formerly. A storm with 50 mm. of rainfall, lasting one hour, may readily apply to an area of 1 sq. meter an amount of energy equivalent to that required to lift a cubic meter of the soil 50 cm above its natural position (8).

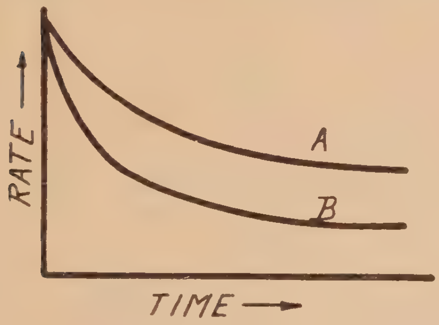
The effect of a single drop has been well portrayed by the stroboscopic photographs of Edgerton in which the explosion and resulting crater in dry soil or the miniature geyser resulting from the impact upon a soil covered with water are clearly seen (8). It must be emphasized that these are the results that are found on bare soil unprotected by a vegetal cover. A vegetal canopy above the soil, however, may largely absorb the energy of rainfall and prevent any serious loss of permeability at the surface. There is thus a marked difference between what happens on a cultivated field having a sparse vegetal cover and one which has a dense cover. The difference is reflected not only in resulting infiltration but also in the amount of erosion.

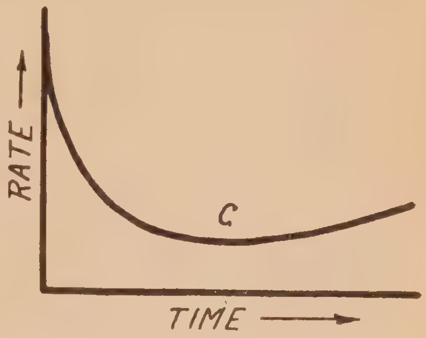
Another factor causing loss of permeability is the slaking of soil aggregates in the presence of water. Those crumbs that are not stable in water readily « melt » down to single grains. The single grains resulting either from rain impact or the slaking of unstable soil crumbs or those discrete particles already present in the surface, are carried by infiltrating water within the soil pores only to lodge at some constricted point within the soil mass. Such migration of fine particles is in effect a natural process of elutriation that probably accounts to a considerable degree for the more dense horizon commonly found beneath the soil surface in humid areas. It occurs in virgin as well in cultivated soils, the difference being primarly one of degree.

Other influences that decrease permeability include the swelling of clays and colloids upon the imbibition of water, an effect, that varies considerably in magnitude in different soils. On irrigated areas base exchange of a character to adversely affect permeability often occurs. Also, it is necessary to note that in extreme circumstances the process of erosion may remove a large portion or all of the surface soil and leave at the surface a relatively dense horizon that formerly lay beneath the surface. Various other influences that commonly modify permeability are the withdrawal of moisture from the soil by vegetation, seasonal changes in temperature, and, of course, the freezing of a wet soil mass. Compacting of freshly irrigated soils by farm tractors has also been reported as inhibiting the permeability of some soils.

Soil Permeability Dynamic and Subject to Modification

The permeability of the soil is thus seen to be dynamic rather than static. Not only do changes occur from season to season and year to year, but also within the duration of a single storm. Typically, the rate of infiltration declines during a storm as shown by curves A and B.





Curves A and B represent the change in rate of infiltration and the range in rate for a single soil in two different states of surface condition. Curve A is that of the soil when in favorable structure as a result of preservation of a normal amount of soil organic matter and good management of cropping practices. Curve B is that of the same soil subjected to intensive cropping to intertilled crops with resulting loss of organic matter and state of aggregation.

Curve C represents an unusual condition in which the rate declines and then gradually and slowly rises. It has been repeatedly found on a deep soil lying high above the water table. The subhorizons are sandy and permeable, and the increase in rate apparently occurs after moisture films have been formed around the soil particles of the subsoil. Among many hundreds of rate curves of different soils that have come to the writer's attention, this is the only soil behaving in this manner.

Practical Application of the infiltration Concept

Within the past five years much progress has been made in the practical application of the infiltration concept. The use of the « runoff coefficient » is gradually being replaced by more logical methods of computing runoff, practically in those instances where adequate infiltration data are available. As Cook (5) has pointed out, however, many problems remain to be solved. At the same time he has shown how infiltration indices may be used and that for a single soil-vegetal cover complex they do have a comprehensible physical meaning. The use of infiltration rate curves for estimating maximum runoff by drawing the predetermined infiltration curve on the intensity-time graph for the given rain and computing the volume of runoff from the area between the two curves after making suitable allowance for surface storage is shown to be a relatively simple process. Various difficulties are encountered particularly when the conditions (of soil moisture for example) under which the infiltration rate curve was derived do not correspond exactly to the conditions under which it is desired to apply the curve. This problem hinges largely upon the adequacy of the basic data, although even here some fairly satisfactory procedures for adjustment from one condition to the other have been developed.

Horner and Jens (14) have been particularly successful in the application of the concept to the design of urban storm sewers and of drainage systems for airports. Similar methods have been used by Hicks in investigations of drainage areas of different sizes and types in the city of Los Angeles, California (11).

The method for computing maximum runoff has also been used by Cook, Horner, Horton, Sherman and others in developing suitable flood control works and land use practices. In certain phases of these developments, which involve estimates of both frequent and infrequent flood water discharges, various additional devices for refining and applying basic data are employed. Throughout all of such work it has been found that the infiltration concept is useful, and whenever basic data are available it is superior to using a « runoff coefficient. »

The infiltration concept has application also in irrigation practice since areas having either very high rates or very low rates usually require specialized treatment or unusual irrigation practices. Low permeability is particularly common in regions where alkali salts prevail either in the soil or the irrigation water. However, any specific situation depends upon the chemical nature of the base exchange that occurs during infiltration. Areas having excessively low rates often necessitate precise leveling of the land and the erection of dikes so that water may be applied under head. Areas having abnormally high rates of infiltration, on the other hand,cannot be irrigated satisfactorily by the furrow method except over very short distances. On such areas, large depths at the head gates are required if the water is to flow the normal lenght of run. The depths required under these conditions are often such as to cause severe erosion (20).

The drainage of land by tile has been found, in at least one large area, to involve essentially the same controlling factors as does the infiltration process. In northwest Ohio, following successful tile drainage ,intensive cultivation resulted in (a)loss of soil organic matter, (b) loss of aggregated structure at the soil surface, (c) low permeability at the soil surface, and (d) failure of the tile system to function properly (3). Obviously, both infiltration and drainage (2) depend upon soil permeability, but it is interesting to note controlling effect of surface permeability in this case upon subsurface drainage.

The relations between infiltration and accretions to the ground water are perhaps considerably more complex than generally realized and clearly call for more researc before they are fully understood. In numerous areas where groundwater depletion has become serious and proposals for remedial measures are made, infiltration among other possible related factors receives condideration (31). In the over-simplified case of a completely permeable profile above the water table, it may be supposed *a priori* that increase in infiltration will be accompanied by greater accretion to ground water. However, even in such a highly simplified instance, one must recognize that all horizons within the profile are not equally permeable and that some unknown quantity of the downward-moving water may be diverted laterally above a less permeable stratum.

Such an oversimplified case becomes more complex when it occurs beneath a mantle of vegetation. Then the vegetation has first opportunity for the withdrawal of water and, within limits, the amount of withdrawal increases as infiltration increases because of normal plant response to available moisture.

An example from an actual case may be cited. The study is a preliminary one and covers only two years (12). It is cited for want of other evidence and must be regarded merely as indicative of the problem until further confirmation is obtained.

On six small watersheds of uniform soil, (three growing bluegrass and three growing corn) rainfall, runoff, infiltration, and the movement of water through the upper 200 cm of the profile were determined. The amount of evapo-transpiration was computed as accretions of water to the root zone minus accretions to depths below root zone. The data in brief show :

1) greater infiltration under bluegrass than under corn;

2) greater evapo-transpiration from grass than corn;

3) greater accretions of water at points below root zone beneath the corn than beneath the grass;

4) the amount of water moving below root zone as measured independently by supplementary equipment was highly and significantly correlated (negatively) with the amount of evapo-transpiration.

The point to be made, of course, is that land use practices that increase infiltration do not necessarily add to ground water. It is important to determine such relationships under various climates, on soils of differing permeabilities and of varying land usages and treatments.

Water spreading for underground storage as practiced in southern California represents an outstanding example of the relations between infiltration and ground water. While the published report of Mitchelson and Muckel (22) well describes the procedures employed in the general vicinity of the San Gabriel and Santa Ana Rivers, the practice has now extended northward into the San Joaquin Valley, where new problems have been encountered and new procedures tested.

The problem essentially involves the temporary storage of irrigation water which seasonally occurs in quantities above immediate needs. The water is spread on filtration beds and allowed to percolate to the underlying aquifer, from which it is subsequently pumped to the surface and applied to irrigated lands. On the sands and gravel beds where the practice was developed, rates of penetration of 30-90 cm, depth per day continuing over extended periods are readily achieved. In the San Joaquin Valley area,however, the soil is of finer texture and lower permeability and there is a tendency for the rate of infiltration to decline after a period of continuous operations. Mitchelson, Muckel, and others are developing various treatments of the surface soil that permit the recovery of the higher rates, although at this date the most effective treatment is not definitely established.

It was found under their conditions that the level of the water table has no effect on the infiltration rate until it comes in actual contact with the saturated soil a few centimeters below the land surface. Any effects of temperature change are so small as to be overshadowed by other more important factors. Possible effects of silt-laden water upon rate of infiltration are obviated by using only clear water upon diked spreading beds.

The rates of infiltration, when plotted to the same scale as hydrographs of observation wells in the underlying aquifer, show the marked response of the water table to added water from above.

Need of further Research

The technician who works in this subject-matter field will recognize the numerous places wherein existing knowledge is inadequate. Existing deficiencies in our understanding of certain parts of the infiltration process greatly limit the practical application of the concept as it relates to soil and plant management and particularly to the prediction of surface runoff for the many different combinations of climates, soils, and plants.

Broadly, the reserach needs fall in several categories :

1. A quantitative evaluation of changes that occur in the soil during a storm, between storms, and from season to season is necessary before general predictions of runoff can be made. This demands a better understanding of the fundamental causes for changing permeability.

2. The facts of the hydrologic cycle being interrelated — the magnitude of one affecting that of another — the practical consequences of infiltration cannot be satisfactorily understood from a knowledge of it alone. There is need, therefore, to determine the relations between infiltration and other phases of the hydrologic

cycle — particularly evapo-transpiration and ground water recharge and discharge — for different climates, soils, and vegetal covers.

3. It also must be recognized that existing techniques for the study of infiltration and its primary modifying factors are far from satisfactory. Studies of « undisturbed » cores are at best studies of a part of the whole and typically a very small part. The very introduction of an instrument into a soil mass may be the means of introducing a new and undesired variable in an investigation.

Radioactive isotopes are now available and offer possibilities of the development of new techniques in this field in the same manner that has been possible in plant and animal physiology. Opportunities of similar kind in studying the movement of fluids through such permeable media as soils should be explored rapidly.

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5)

NOTE SUR LA PERCOLATION PROFONDE DES PRECIPITATIONS D'EAU DANS LES MASSIFS CALCAIRES EN TUNISIE

Présentée par la Direction des Travaux Publics

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1. — INTRODUCTION

En général l'étude de la percolation profonde des précipitations d'eau dans le sous-sol est malaisée du fait de l'extrême complexité de ce phénomène.

Néanmoins elle a pu être tentée en Tunisie dans quelques cas particuliers simples, notamment dans le cas de massifs calcaires de dimension limitée et dont on connaît tous les exutoires d'eau,

Ainsi trois études ont pu être faites concernant la percolation profonde des précipitations dans les calcaires du Djebel Chennata près de Bizerte, du Djebel Bent Saidane près de Pont du Fahs et du Djebel Zaghouan près de Tunis.

Ces réserves aquifères sont exploitées pour l'alimentation en eau des villes de Tunis et de Bizerte.

Dans notre travail nous nous sommes basés sur les relevés hebdomadaires du débit des sources et du niveau d'eau dans les forages faits par la Compagnie des Eaux de Tunis et la Subdivision Spéciale des Eaux de Bizerte. Nous nous sommes aussi servi de renseignements pluviométriques et de mesures d'évaporation faits par le Service Météorologique de la Tunisie.

2. — MASSIFS CALCAIRES ETUDIES

I. — Djebel Chennata

Le Djebel Chennata se trouve situé 13 kilomètres environ à l'Ouest de la ville de Bizerte.

Cette montagne est constituée essentiellement par un monoclinal de calcaire Campanien de 3 kms de longueur. Elle est limitée aux deux extrémités par des failles transversales étanches. Les calcaires y sont légèrement marneux et comportent des interstratifications marneuses.

L'épaisseur du Campanien est évaluée à 150 mètres environ, le pendage étant environ de 20° vers le N.W. (voir fig. 1).

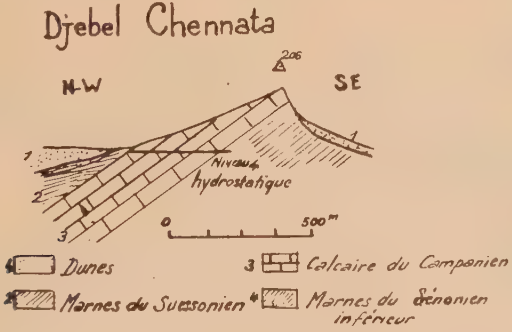

Fig. 1.

Les marnes du Suessonien et celles du Sénonien inférieur forment le toit et le mur du Campanien qui recèle le niveau d'eau exploité pour la ville de Bizerte.

La géologie de cette montagne a été bien étudiée par M. Solignac (1).

Une galerie de captage avec serrement fait ponction dans ces calcaires et permet de drainer une tranche de calcaire de 30 mètres de haut.

Un forage exécuté en amont de la galerie permet de suivre constamment les mouvements du niveau hydrostatique dans les calcaires aquifères, se produisant à la suite des pluies ou de l'exploitation.

L'alimentation des calcaires se fait uniquement par infiltration des eaux météoriques.

Nous évaluons à 1,8 kilomètre-carré, la surface du périmètre d'alimentation de ces calcaires.

Le quart de cette surface est couvert de dunes plus ou moins consolidées.

Le reste est couvert de brousse très dégradée formée par l'association végétale de lentisques et d'oliviers (Oleo-lentiscetum de R. Maire) (2).

La pluviométrie de cette région peut être connue grâce au poste météorologique de Bouzaréa, à 3 kms environ au Sud du Djebel Chennata; il fonctionne depuis une dizaine d'années.

La hauteur moyenne des précipitations, d'après le poste de Bizerte, pour la période de 1901-1940 était de 636 millimètres par an. La moyenne mensuelle étant de :

Septembre	33	Mars	55
Octobre	69	Avril	41
Novembre	88	Mai	21
Décembre	116	Juin	13
Janvier	105	Juillet	9
Février	81	Août	5

La pluviométrie est relativement régulière. Les valeurs extrêmes étant : 390 m/m et 931 m/m par an.

L'intensité d'évaporation déduite de 10 années d'observation à l'évaporimètre Piche, faite au même poste, serait de 1279 millimètres par an, la moyenne mensuelle étant de :

Septembre	123	Mars	83
Octobre	102	Avril	90
Novembre	75	Mai	123
Décembre	72	Juin	147
Janvier	68	Juillet	174
Février	69	Août	153

II. — Djebel Bent Saïdane

Le massif de Bent Saïdane fait partie de la chaîne dorsale Tunisienne. Il est situé 20 kms, environ au Sud de la ville de Pont du Fahs, entre le Djebel Zaghouan et le Djebel Fkirine.

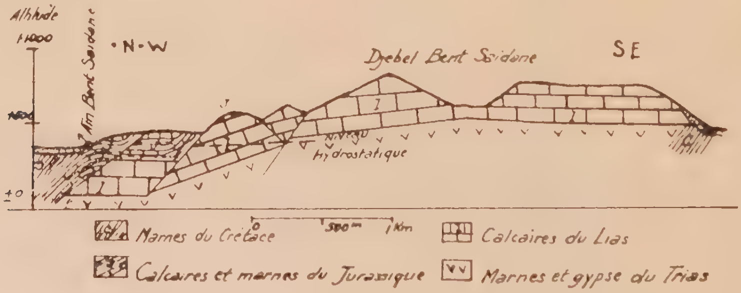

FIG. 2.

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(1), (2), les chiffres renvoient à la bibliographie.

Il est formé essentiellement par des monoclinaux de calcaire du Lias de 250 m d'épaisseur limités de toute part par des failles ayant joué les unes par rapport aux autres (voir fig. 2).

Les calcaires du Lias sont subcristallins ou siliceux. Ils sont généralement en gros bancs.

Les marnes du Trias, du Crétacé et vraisemblablement de l'infra-Lias forment le mur; les marnocalcaires du Crétacé et du Jurassique forment le toit du Lias.

Ces calcaires recèlent un niveau aquifère important exploité pour les besoins en eau de la ville de Tunis.

Deux sources : l'Aïn Djouggar et l'Aïn Bent Saïdane captées à cet effet, émergent sur le versant N.W. de la montagne.

Nous évaluons la surface des affleurements du Lias à 16,4 kilomètres-carrés. Ces affleurements sont couverts par la formation Pinetum halepensis dégradée par le pacage et les incendies réduite à l'état de broussaille clairsemée.

On y trouve surtout le lentisque (Pistacia lentiscus), le chêne vert (Quercus ulex) le romarin (Romarinus officinalis) les cistes (Cistus) et les plantes herbacées des clairières, à la place du pin (Pinus halepensis).

La hauteur moyenne des précipitations atmosphériques y est de 508 millimètres par an pour la période de 1901 à 1940, d'après les observations faites au poste météorologique de Djouggar qui se trouve au pied du versant Nord du Massif de Bent Saïdane.

La moyenne mensuelle pour la même période est de :

Septembre	46	Mars	58
Octobre	54	Avril	44
Novembre	53	Mai	30
Décembre	57	Juin	20
Janvier	65	Juillet	6
Février	58	Août	15

Les extrêmes sont de 321 mm. et 918 mm. par an.

L'intensité de l'évaporation y est moins bien connue. Mais on peut supposer qu'elle est comparable à celle observée au barrage de l'Oued Kebir, à l'aide d'un bac de 3 m de côté, à 18 kms à l'Ouest du massif étudié.

Cette évaporation atteint 1400 millimètres par an, la moyenne mensuelle est de :

Septembre	144	Mars	59
Octobre	103	Avril	118
Novembre	67	Mai	141
Décembre	46	Juin	213
Janvier	45	Juillet	212
Février	41	Août	211

III. — Djebel Zaghouan

Le massif calcaire de Zaghouan fait partie de la chaîne dorsale Tunisienne. Il se trouve 50 kms environ à vol d'oiseau au Sud de la ville de Tunis. C'est l'une des plus hautes montagnes de la Tunisie.

La masse principale de cette montagne est formée de monoclinaux liasiques calcaires déracinés et chevauchant les uns sur les autres.

Ces monoclinaux sont limités de toute part par des failles et entourés de marnocalcaires et de marnes du Crétacé et de l'Eocène. Ils reposent sur les marnes irisées et les grès en plaquettes du Trias.

La stratigraphie détaillée du Djebel Zaghouan a été faite par M. Solignac (1). La tectonique a été élucidée par M. G. Castany (2) (voir fig. 3).

Les calcaires du Lias sont subcristallins en gros bancs ou massifs de 350 mètres d'épaisseur totale. Ils renferment un niveau d'eau abondant.

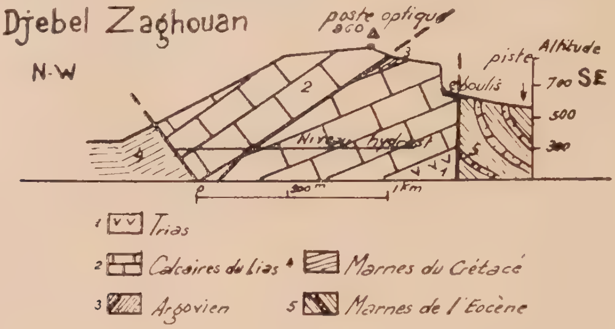
Trois sources captées pour la ville de Tunis lui sont attribuables : Aïn el Guelb, Aïn Ayed et Nymphée.

Le débit de ces sources est relevé systématiquement par la Compagnie des Eaux de Tunis depuis 40 ans au moins, avec une précision très inégale, il est vrai.

L'alimentation du niveau d'eau des calcaires du Lias doit se faire de manière habituelle par infiltration des eaux météoriques.

Nous évaluons à 19,6 kilomètres-carrés, la surface du périmètre d'alimentation des calcaires du Djebel Zaghouan.

Ce périmètre est couvert de brousse de *pinetum halepensis* clairsemée et très dégradée par le pacage et les incendies.



D'après M. Castany

FIG. 3.

D'après les observations faites au poste météorologique de Zaghouan (Contrôle Civil), la hauteur moyenne des précipitations serait de 513 millimètres par an pour la période 1901-1940.

La moyenne mensuelle pour la même période étant de :

Septembre	45	Mars	65
Octobre	53	Avril	41
Novembre	57	Mai	28
Décembre	56	Juin	16
Janvier	71	Juillet	4
Février	62	Août	13

Les extrêmes sont de 259 m/m et 923 m/m par an.

L'intensité de l'évaporation y est moins bien connue. Nous admettrons, comme pour le massif de Bent Saïdane, que l'évaporation y est peu différente de celle observée au barrage de l'Oued Kebir, à une distance de 35 kms.

3. — ETUDE DE LA PERCOLATION

I. — Principe

Notre étude était basée sur l'évaluation des apports d'eau au sous-sol, en tenant compte des fluctuations observables du plan d'eau libre dans des puits témoins, ou des variations du débit des sources.

Le calcul des apports d'eau a été fait en assimilant la roche aquifère à un réservoir d'eau pluviale qui se remplit et se vide alternativement.

Notre travail était fait en trois stades successifs :

A) Recherche de la relation entre l'altitude du niveau hydrostatique (ou le débit des sources) et la réserve d'eau accumulée en sous-sol.

B) Evaluation des apports d'eau successifs en tenant compte des fluctuations observées du niveau hydrostatique ou des variations du débit des sources.

C) Recherche de la relation existante entre la hauteur des précipitations observées et le volume des apports d'eau calculés.

II. — Réserves d'eau

A. — *Djebel Chenata*

On a cherché la relation existant entre la réserve d'eau exploitable par le captage et l'altitude du plan d'eau libre dans le puits témoin.

Cette évaluation a été faite en calculant les volumes d'eau puisés dans le sol en période sèche correspondant aux abaissements successifs du plan d'eau dans le puits de 2,5 mètres en 2,5 mètres.

On a établi ensuite, la relation entre l'altitude du plan d'eau et la réserve susceptible d'être puisée dans le sol par addition des volumes d'eau correspondant aux différents abaissements successifs observés.

Cette façon de procéder n'est pas sans reproches. Mais c'est la seule possible, et elle suffit dans les conditions actuelles d'exploitation.

B. — *Massif de Bent Saïdane et Djebel Zaghouan*

L'évaluation des réserves d'eau a été faite en se basant sur la courbe de débit des sources en fonction du temps en régime non influencé.

Cette courbe à été obtenue par superposition avec translation suivant l'axe des temps de plusieurs tronçons de courbes réelles observées correspondant aux différentes périodes sèches successives.

En partant de cette courbe on a établi par intégration le graphique des volumes d'eau susceptibles d'être puisés dans le sol par les sources en fonction du débit initial de l'une d'elles : Aïn Djouggar ou Nymphée.

On a trouvé ainsi que la réserve exploitable est pour :

Djebel Chennata : 470.000 m^3 pour une variation de niveau hydrostatique de 29 mètres.

Djebel Zaghouan : 2.700.000 m^3 pour la variation de débit des sources Nymphée et Aïn Ayed, comprise entre 15.000 et 4.500 m^3/jour.

Massif de Bent Saïdane : 4.000.000 m^3 pour les débits de la source Aïn Djouggar, compris entre 14.000 et 3.200 m^3/jour.

III. — Apports d'eau

L'évaluation des apports d'eau souterraine a été faite en se basant sur les augmentations ou les diminutions successives de réserve d'eau, compte tenu du volume puisé dans le sol.

Elle a été calculée suivant l'expression :

$$a = (R_2 - R_1) + v$$

avec :

a — apport d'eau

R_1 et R_2 — réserve d'eau correspondant au débit des sources ou à l'altitude du plan d'eau du puits témoin observée respectivement au début et à la fin de la période envisagée.

v — volume d'eau puisé dans le sol pendant cette période.

On a trouvé ainsi que l'apport annuel médian était :

Chennata	9 ans d'observ.	580.000 m^3
Zaghouan, ...	18 ans d'observ.	3.200.000 m^3
Bent Saïdane, ...	16 ans d'observ.	1.450.000 m^3

L'apport minimum étant égal ou inférieur à :

Chennata, ...	346.000 m^3	1 fois en 9 ans
Zaghouan, ...	1.800.000 m^3	1 fois en 6 ans
Bent Saïdane, ..	0 m^3	1 fois en 8 ans

L'apport maximum étant :

Chennata : plus de	840.000 m^3	1 fois sur 9 ans
Zaghouan : plus de	5.000.000 m^3	1 fois sur 6 ans
Bent Saïdane : plus de ..	4.900.000 m^3	1 fois sur 8 ans

IV. — Relation entre la pluviométrie et la percolation

A. — *Formules*

La recherche de la relation existant entre la hauteur des précipitations et le volume des apports d'eau a été faite en déterminant la valeur des paramètres Ho, S' et du coefficient α de la formule théorique suivante :

$$A = S' \left[\sum h - Ho - \alpha \sum e\right]$$

avec :

S' — Surface du périmètre d'alimentation effectif.
$\sum h$ — Hauteur cumulée des précipitations depuis le début de la suite pluvieuse — envisagée jusqu'à la fin de la dernière pluie ayant pu saturer le sol.
A — volume d'eau infiltrée dans le sol correspondant aux précipitations h envisagées.
Ho — Hauteur d'eau susceptible d'être retenue par le sol.
α — Coefficient numérique.
$\sum e$ — Hauteur cumulée d'évaporation d'eau « e » correspondant à l'intervalle de temps existant entre la première précipitation de la suite pluvieuse envisagée et la dernière pluie ayant pu saturer le sol.

La formule est basée sur les hypothèses simplificatrices suivantes, qui sont appliquées simultanément.

1e *hypothèse* : La surface du sol (terre végétale) a une capacité de rétention Ho uniforme sur toute l'étendue du bassin d'alimentation du niveau aquifère étudié.

2e *hypothèse* : La percolation d'eau au sous-sol commence seulement après la saturation du sol par Ho millimètres d'eau.

3e *hypothèse* : L'eau retenue par le sol est soumise à l'évaporation par la végétation et la surface du sol (reprise atmosphérique) ; l'eau percolée au sous-sol en étant exempte.

4e *hypothèse* : La quantité d'eau reprise par l'atmosphère est proportionnelle à l'évaporation physique (mesurable par l'évaporimètre Piche ou par bac).

Cette formule est valable seulement pour les valeurs positives de A.

Elle peut être employée pour une pluie isolée ou pour une succession de précipitations suffisamment rapprochées pour qu'il reste constamment de l'eau retenue par le sol malgré la reprise atmosphérique.

Cette dernière condition nous obligera parfois à partager une saison pluvieuse en plusieurs périodes de percolation, le calcul étant à faire séparément pour chacune d'elles.

La figure 4 est un exemple des graphiques utilisés pour la détermination de Ho et α.

B. — *Graphiques*

La recherche de la valeur des paramètres S', Ho et du coefficient α a été faite graphiquement par des approximations successives.

A cet effet on a établi :

1 — Un graphique de pluie cumulée et des apports d'eau.

2 — Plusieurs faisceaux de courbes $y = [\alpha \sum e + C]$ tracées sur du papier calque à la même échelle que le graphique précédent (voir fig. 4), C étant une hauteur de référence.

Une feuille de papier était utilisée pour chaque valeur de « α » convenablement choisie et variant de 0,1 en 0,1.

Sur chaque feuille le faisceau était constitué par des courbes correspondant aux différentes valeurs de C variant de 50 m/m en 50 m/m.

L'évaporation admise $\sum e$ était l'évaporation moyenne de la plus proche station d'observation : Poste de Bizerte Karouba pour Dj. Chennata ; poste de l'Oued Kebir pour Zaghouan et Bent Saïdane.

La recherche de S', Ho et α était faite par tatonnement en superposant au graphique de pluie les différents faisceaux de courbes $y = [\alpha \sum e + C]$ et en envisageant la ligne de pluie comme étant tracée en coordonnées curvilignes « C » et temps « t ».

C. — *Recherche de α et Ho*

Le choix de la valeur de « α » et de Ho était fait en se basant sur les remarques suivantes :

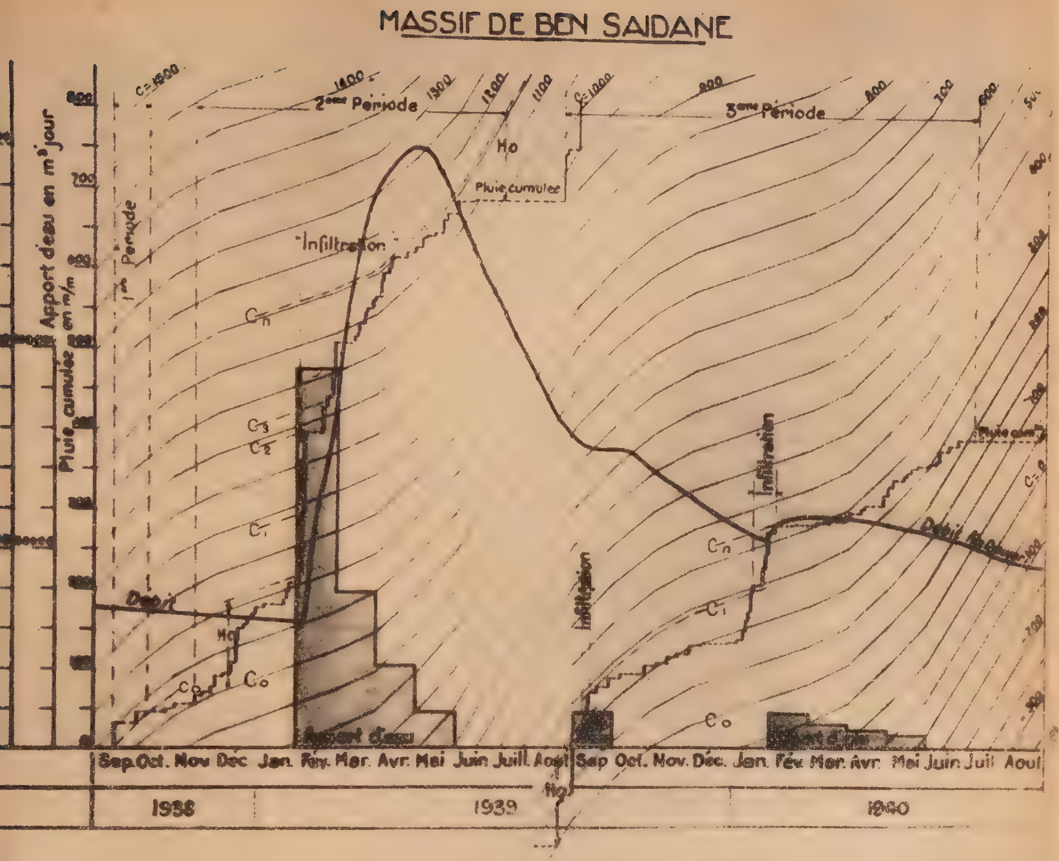


Fig. 4

1e remarque : Pour une valeur convenablement choisie de « α », on doit trouver l'inégalité suivante :

$$Co + Ho < C_1 < C_2 < C_3 \ldots < Cn$$

avec :

Co — Coordonnée du commenceemnt de la première précipitation de la période de percolation envisagée.

C_1, C_2, C_3 ... Cn — Coordonnées de la fin de la 1e, 2e, 3e et ne dernière précipitation ayant saturé le sol.

2e remarque : La coordonnée C de la ligne de pluie passe par son maximum à la fin de la dernière précipitation avec apport d'eau au sous-sol. Autrement dit $Cn = C$ maximum.

3e remarque : Pour la valeur convenablement choisie de α et Ho, on doit avoir :

$$S' = -\frac{A}{Cn - Ho - Co} = \text{constant}$$

D. — *Détermination de* S'

L'expression $Cn - Ho - Co$ qui entre dans la formule précédente représente la hauteur d'eau « *p* » effectivement percolée au sous-sol.

Elle se relève aisément sur nos graphiques.

Nous avons donc admis pour S' la valeur donnée par l'expression :

$$S' = \frac{\sum A}{\sum p}$$

avec :
$\sum A$ — la somme des apports d'eau calculés pour toute la durée des observations.
$\sum p$ — la somme des percolations pendant la même période.

Dans ce calcul nous n'avons pas tenu compte des années avec des apports d'eau dépassant la capacité d'emmagasinement de la roche aquifère : l'excédent d'eau était alors évacué par des émergences temporaires et le ruissellement. Le calcul est alors en défaut.

E. — *Valeurs admises de* Ho, S′, *et* « α »

En procédant ainsi nous avons trouvé les valeurs suivantes :

Chennata ..... : H*o* = 60 m/m $\alpha = 0{,}5$ S′ = 1,8 km²
Zaghouan .... : H*o* = 85 m/m $\alpha = 0{,}4$ S′ = 20 km²
Bent Saïdane : H*o* = 120 m/m $\alpha = 0{,}4$ S′ = 17 km²

Il est à remarquer que :

1 — La valeur la plus faible du paramètre H*o* trouvé, correspond aux calcaires relativement tendres d'une région à pluviométrie régulière. La valeur la plus forte de H*o* correspond aux roches dures d'un massif à pluviométrie capricieuse.

2 — La superficie effective S′ du bassin d'alimentation est très voisine de celle du périmètre d'affleurement des roches aquifères « S ». En effet on a :

Chennata .... S = 1,8 km² S′ = 1,8
Zaghouan .... S = 19,6 S′ = 20
Bent Saïdane ...... S = 16,4 S′ = 17

Nous déduisons de la quasi égalité entre S et S′ que le ruissellement y était bien négligeable par rapport à l'infiltration au cours des années retenues.

3 — La valeur du coefficient moyen de percolation profonde K correspondant aux valeurs de H*o*, α et S′ admises, donnée par l'expression :

$$K = \frac{S' \sum p}{S \sum h}$$

serait de :

Djebel Chennata .... K = 0,52
Djebel Zaghouan .... K = 0,33
Massif de Bent Saïdane ...... K = 0,26

La hauteur moyenne des précipitations étant pour la période envisagée de :

Djebel Chennata .... 787 millimètres
Djebel Zaghouan .... 490 millimètres
Massif de Bent Saïdane ...... 503 millimètres

F. — *Comparaison des résultats de calculs de la percolation* « *p* » *avec les observations faites* :

L'examen du graphique établi avec les valeurs de « α » admises par nous, nous a montré que :

1 — L'instant du commencement des percolations d'eau au sous-sol, observé et déterminé par le calcul concorde, à l'écart correspondant aux irrégularités d'évaporation près.

2 — Le volume annuel des apports d'eau observé et calculé concorde de manière satisfaisante.

Les écarts entre les valeurs observées et calculées de l'apport annuel exprimé en millimètres d'infiltration sont indiqués dans le tableau suivant :

Massif	H*o* admis m/m	écart médian m/m	écart maximum m/m	Percolation médiane « *p* » m/m	Nombre d'années comparées
Chennata	60	12	42	445	9
Zaghouan	85	20	77	163	15
Bent Saïdane	120	24	65	90	12

Il est à remarquer que l'écart constaté entre les résultats de calcul et d'observation est d'autant plus élevé que H*o* est plus grand.

Nous attribuons les écarts aux irrégularités d'évaporation.

Mais actuellement on ne peut pas espérer obtenir de meilleurs accords entre l'observation et le calcul étant donné l'absence de mesures précises d'évaporation et l'approximation des hypothèses admises.

Nous nous proposons donc d'installer des évaporimètres dans les massifs calcaires étudiés et de reprendre les observations pour préciser la relation existant entre les précipitations atmosphériques et l'infiltration d'eau au sous-sol.

Mais dès maintenant, nous estimons que les résultats obtenus nous permettent de suivre les fluctuations des réserves aquifères étudiées, de régler leur exploitation et de prévoir l'importance des apports d'eau utilisables dans des massifs placés dans des conditions analogues.

Tunis, le 13 mars 1948.

L'Ingénieur en Chef du Groupe N.A.R.,
J. TIXERONT.

Le Géologue Chef du B.I.R.H.
E. BERKALOFF.

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6) INFILTRATION (ABSORPTION BY THE SOIL), FACTORS WHICH DETERMINE THE DEGREE OF INFILTRATION OF THE RAINFALL, TAKING INTO ACCOUNT THE EFFECT OF PLANTS, THE CONFIGURATION OF THE SOIL, ETC. AND THE IMPORTANCE OF THESE INFLUENCES

Report by Dr. **R. K. SCHOFIELD** (Great Britain)

Close relationships of transpiration by plants to evaporation from open water

Recent work by H. L. Penman at Rothamsted Experimental Station demonstrates the close relationships of transpiration by plants to evaporation from open water. Use was made of cylindrical tanks 6 ft. deep and 2 ft. 6 in. in diameter which had been sunk into the earth so that their rime project about 1 in. above the surrounding soil. Ten of these have been connected in pairs. One of each pair contains soil to the level of the surrounding soil, while the other is covered and acts as a reservoir. Three of the soil tanks carry grass. The water table in these is maintained at any desired depths (e.g. 10 inc, 16 in. and 24 in.) by observing the depth of water in reservoir tanks and adding or removing measured quantities of water from them when necessary. Two soil tanks are kept bare. In one of these the water table is held at 5 in. and the soil surface is always moist. One cylinder (not connected to another) is kept filled with water freely exposed.

For convenience of experimentation these cylinders had been installed round half the circumference of a brick-lined pit 20 ft. in diameter and 8 ft. deep. Their exposure, particularly to heating by solar radiation, is not quite normal, as is shown by the fact that on sunny summer days the evaporation from another water tank exposing the same area sunk in the ground well away from the pit is some 25% less than from the water tank on the edge of the pit and that there is a corresponding difference in surface temperature. Moreover, each of the tanks round the pit has a slightly different aspect, and some allowance may have to be made for this. Nevertheless it is clear that while evaporation from the damp soil is nearly always a little greater than that from the grass and a little less than from the open water, all three respond in essentially the same way to weather conditions.

Transpiration is not closely related to rate of growth

In the spring of 1946, fertilizers were applied to one of the grass tanks but not to the other two. During the ensuing summer the water-table in all three was held at 24 in. The yield of cut garss from the fertilized tank was three times that from either of the others yet the evaporation was practically the same from all three, showing that growth-rate is not directly connected with transpiration.

Evaporation from grass with ample water supply.

Approximate factors connecting the evaporation from subirrigated grass with the evaporation from open water have been obtained at Rothamsted. These have been used to interpret records of a grass drain-gauge situated beside an evaporation tank at the Harrogate Water Works. Here on Blubberhouses Moor, 549 ft. above O.D., the rainfall distribution and evaporation are such that the grass is scarcely ever short of water. The table shows how an estimate of 15 in. mean annual evaporation from the grass has been obtained from the mean monthly evaporation from the open water tank. This is in satisfactory agreement with the excess of mean annual rainfall over mean annual drainage from the grass covered gauge which is 15.2 in.

	E_0 in.	f	E_0f in.
January	0.41	0.6	0.25
February	0.45	0.6	0.27
March	0.98	0.7	0.68
April	1.78	0.7	1.25
May	2.71	0.8	2.18
June	3.28	0.8	2.62
July	3.38	0.8	2.70
August	2.73	0.8	2.17
September	1.94	0.7	1.36
October	1.25	0.7	0.88
November	0.62	0.6	0.38
December	0.43	0.6	0.26
Year	19.96		15.00

E_0 = mean evaporation from open water.
E_0f = estimated mean evaporation from grass.

	inches
Mean annual rainfall	30.64
Mean annual drainage	15.45
Difference	15.19

Evaluation of Soil Moisture Deficits at Cambridge.

In many places in Great Britain there are often periods in summer when grass growing under normal field conditions or in a drain-gauge does not evaporate as much water as subirrigated grass. Records relating to such conditions have been obtained by H. H. Nicholson for grassland on the Cambridge University Farm. Observations have been made of the times when the mole-drainage system discharges. From these records it is possible to determine the last date in spring when there was no deficit for soil moisture and then to find when next the accumulated rainfall had again overtaken the accumulated evaporation as shown by the free running of the drains. When the summer's evaporation is compared with the estimated evaporation from an open water surface it becomes clear that some allowance must be made for restriction of transpiration due to dryness of the soil. Penman has done this by assuming that evaporation was equal to that from subirrigated grass until a deficit had been reached, after which a restriction was imposed. A value of c was found by trial from each year's record to fit the observed time of the restart of drainage.

Direct observations of soil moisture

Values obtained in this way for the soil moisture deficits are shown graphically in the figure. In 1933 and 1934 it is possible to check the estimated deficits against the results of soil moisture determinations made by Nicholson and his students. Even though the soil of this grassland is exceptionally uniform the sampling error in such a direct determination of soil moisture deficit may easily be in error by one inch.

A technique that reduces the sampling error due to soil irregularity is being explored at Rothamsted. It has been found that when the soil moisture deficit in the local soil can be taken as zero, the soil water has a « suction » of about 50 cm of water. Clods taken when a moisture deficit exists are allowed to take water against a suction of 50 cm until they attain constant weight. If clods taken from a series of depths down the profile are studied in this way an estimate can be made of the total soil moisture deficit and its distribution with depth.

For many it is necessary to know the total soil moisture deficit but not its distribution with depth In such cases the indirect approach based on estimated evaporation appears likely to give more accurate results and is certainly less laborious.

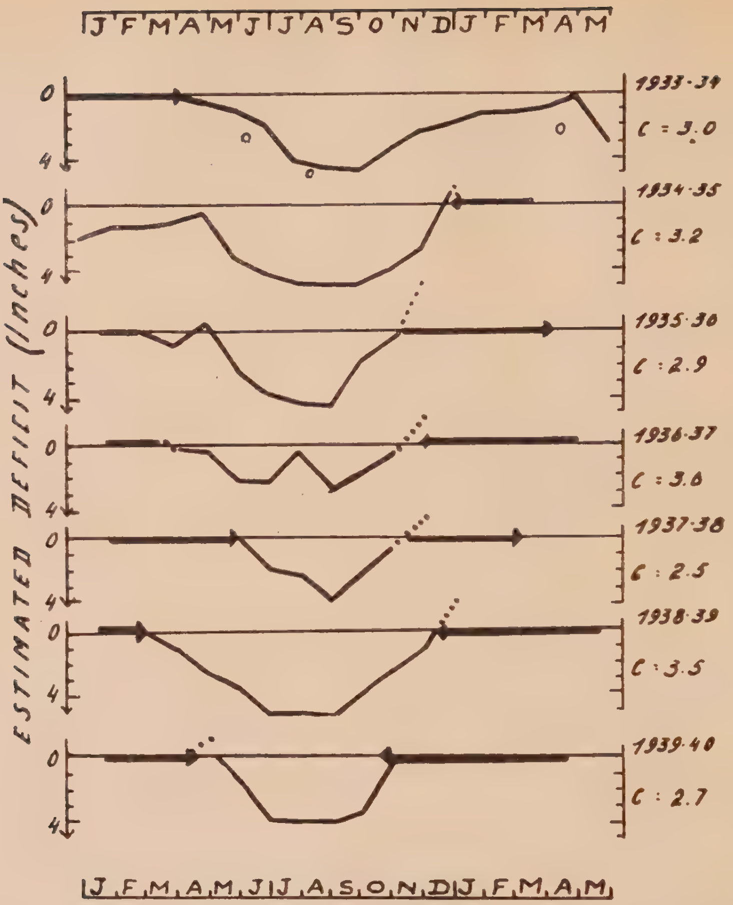

Fig. 1. Estimated and observed soil moisture deficits in grassland at Cambridge. Thick horizontal lines show when the field drains were running. Open circles show values of deficit measured by soil sampling (1933-34).

Estimation of Evaporation from Meteorological Data

Until more measurements have been made of evaporation from vegetation, and more well-site evaporation tanks are installed, it will be necessarry to supplement direct observations by estimates based on meteorological data. It may be remarked in passing that observed evaporation from open water should not be underrated as an index of weather and climate. Of the meteorological factors that determine evaporation, the incident solar ratiation is certainly the most important. The turbulence. temperature and humidity of the air, and the exchange of thermal radiation all help to determine the proportion of the incident energy consumed as latent heat of vaporization. Penman has written down the equations for evaporation, one based on the energy balance, and the other on Rohwer's equation for the dependence of vapour transfer on wind speed. The first involves surface temperature explicitly while the second involves the vapour pressure at the surface which, in the case of a free water surface, is determined by surface temperature. By combining the two equations Penman eliminates surface temperature and can make a fairly accurate estimate of the evaporation from open water giving the latitude and mean seasonal values of hours of sunshine, wind speed, air temperature and dew point.

Interpretation of water-table fluctuations

W. V. Lewis (1943) has studied the fluctuations in level of the Breckland Meers and shown that they reflect the rise and fall of the water-table in the surrounding country. He finds a parallelism between his observations and the records of drain gauges in S. E. England covering the same period. He develops a formula for estimating the annual drainage from the surrounding monthly rainfalls.

E. C. Childs (1943-1947) has worked out in great detail the analogy between equipotentials and streamlines in electrical cunductivity and those that would be expected in idealised cases of drainage to a water-table.

Rate of absorption of water by dry soil

H. H. Nicholson and E. C. Childs (1936) and E. C. Childs (1943) have studied the performance of mole drains in the grassland on the clay soil of the Cambridge University Farm. They obtained evidence, particularly in summer, that sudden rains can reach the drains and cause discharge before the absorptive capacity of the soil has been fully satisfied. Childs estimated that if the rain lasts long enough, absorption may continue at a decreasing rate and only become negligible in 4 hours. Use was made of a recording raingauge and instruments specially designed to record instantaneous rates of flow from drain outfalls.

H. L. Penman and R. K. Schofield (1941) observed the same phenomen in their analysis of the continuous records of drainage from the Rothamsted draingauges (bare undisturbed clay soil containing flints). On one occasion drainage occurred when the soil, given time, could still absorb 0.5 in. of rain. In both these soils the water not absorbed passes rapidly downwards through cracks.

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COMMISSION DES EAUX SOUTERRAINES

Rapports relatifs à la deuxième question :
VARIATIONS DES NIVEAUX DE L'EAU

1) THE EFFECT OF THE YEARLY FLUCTUATIONS IN RAINFALL ON THE FLOW OF GROUND-WATER FROM AN EXTENDED AREA OF RECHARGE

Investigations into the fluctuations in the natural discharge of ground-water from the Veluwe-district in the Netherlands

Paper presented by the Netherlands State Institute for Water Supply.

I. Abstract

The natural discharge of ground-water from any area of recharge undergoes the influence of the seasons, the effect of the succession of alternate dry and wet periodes. As appears from a mathematical analysis, the effect of the precipitation on the variations in the discharge of ground-water from an area of infinite extent, for the case where the intensity of the precipitation on this area varies as a sine with a period of a year, is in some respect analogous to the tidal fluctuations in a body of ground-water that communicates with open water in which a tidal oscillation, likewise with a period of a year, is maintained. So the formulas for this tidal effect may be used when the influence of varying precipitation on the discharge of ground-water has to be investigated.

A second way for studying the problem is to be found in model-research.

By comparing the results of the analytical investigations with those of the model-research ,an insight on the reliability of the solution may be gained.

In this paper is described how both methods have been used for investigating the variations in the discharge of ground-water from the hilly Veluwe-district in the central Netherlands and the delay in these variations compared to the variations in rainfall.

II. Introduction and summary

Among the important engineering works which have been carried out in the Netherlands in the last decades may be classed the damming off and the partial reclamation of the Zuiderzee, the wide inlet of the North Sea which penetrated deep in to the heart of the Netherlands.

The scheme includes the reclamation of four parts of the former Zuiderzee. These four « polders » are numbered from 1 to IV in Fig. 1. The reclamation of polder 1 and II has been finished; the polders III and IV will be tackled in the near future. The remaining portion of the dammed-off brackish sea forms an inland lake, the Yssel-Lake, the water of which gradually became fresh. The Yssel-Lake may meet the ever-increasing demand for drinking-water for the densely populated western provinces of the country. Not everywhere in the lake, however, the water is equally fit for this purpose. It is expected that the best quality will be found in the long but narrow stretch of water which will be left in the south-east corner of the former Zuiderzee between the old coast and the future polder N° 4. After the main influent river, the Eem River, this oblong lake will be called the Eem Lake.

The Eem Lake gets its water from three sources; the polder No 4 drains its excess of water into the lake; the Eem River and several smaller streams empty into it; the ground-water from the bordering mainland, the hilly Veluwe-district flows into the lake.

Now the last-mentioned supply of water is the principal one; especially in the dry season, when the influent rivers have a small discharge and the polder has little or no superfluous water. So it depends largely on the recharge of the lake by the inflowing ground-water whether it can be used as a source capable of providing always a sufficient quantity of potable water of high quality. It is. therefore, of great interest to know the variations in the discharge of ground-water from the

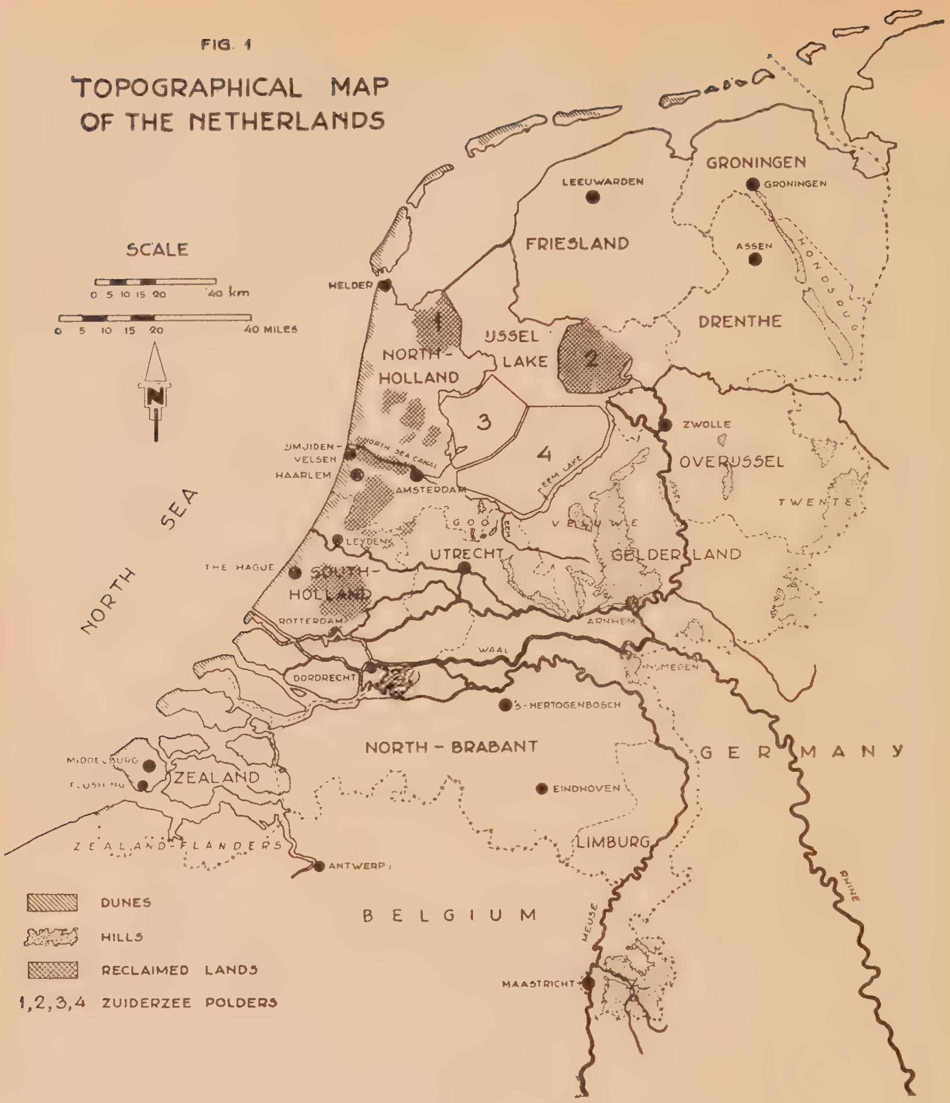

FIG. 1.

Veluwe-district. For this purpose the general problem of the variations in the discharge of ground-water from an extended area was submitted to analysis for the simplified case where the intensity of the precipitation on the area varies as a sine. (Whenever rainfall or precipitation is mentioned in this paper, always rainfall minus evaporation is meant). From this analysis it appeared that there exists an analogy between this phenomenon and the propagation of tidal fluctuations in a body of ground-water, communicating with open water in which the tidal fluctuations is maintained, the period of the tide of course being equal to that of the variations in the precipitation in the first-mentioned problem. In consequence of this analogy formulas governing tidal propagation in ground-water can be used for solving the problem in question. In this way the variations in the discharge of ground-water. from the Veluwe-district could be computed.

In the second place the problem was investigated by means of model-research In a model the body of ground-water of the Veluwe-district was imitated on a small scale and the variations in the discharge at the end of the model were measured, the intensity of the rain supplied in the model varying as a sine with a period of a year.

By comparing the results of both widely differing methods of investigation, viz. the mathematical -analytical one and the model-research, an insight could be gained into the accuracy of the conclusions drawn from the investigation. It appeared that the results of the model-research agreed very well with those of the mathematical analysis, and so both methods of investigation may be accepted as sufficiently reliable.

In the sections to follow will be dealt with successively (1) the hydrological character of the Veluwe-district, (2) the mathematical analysis of the general problem of the varying discharge of ground-water from an area with infinite extent, the intensity of the precipitation over this area varying as a sine, (3) the application of the derived formulas to the case of the Veluwe-discharge, (4) a description of the model used for the investigations and the results of the model-research.

At the end of this paper references to literature on the various subjects are given.

III. The hydrological character of the Veluwe-district

The Veluwe-district is one of the few hilly regions of the Netherlands, the hills however attaining an altitude of no more than from 50 to 100 m above sea-level. The Veluwe is bound in the north by the coast of the former Zuiderzee, in the east by the Yssel river, in the south by the Rhine and in the west by the valley of the Eem River (Fig. 1). The subsoil of practically the whole area enclosed between these four boundaries, down to a depth of more than 200 m below sea-level or about 300 m below land surface, consists of plistocene deposits of coarse sand and graved, for the greater part geologically belonging to the glacial and interglacial periods of the Riss-Würm. These deposits are underlain by formations of Icenian age which are practically impervious to ground-water. Thus a ground-water basin with a depth of nearly 300 m and of great water-carrying capacity is present here. From the latest hydrological investigations the transmissibility of this enormous aquifer appears to amount to some 7500 m^2 per day. (As may be considered well-know, the transmissibility T of an aquifer may be defined as the product of the permeability-constant K of the soil and the thickness *m* of the aquifer).

The ground-water in this Veluwe-aquifer is continuously being replenished by rainfall, varying in the course of the seasons. The rainfall (minus evaporation)

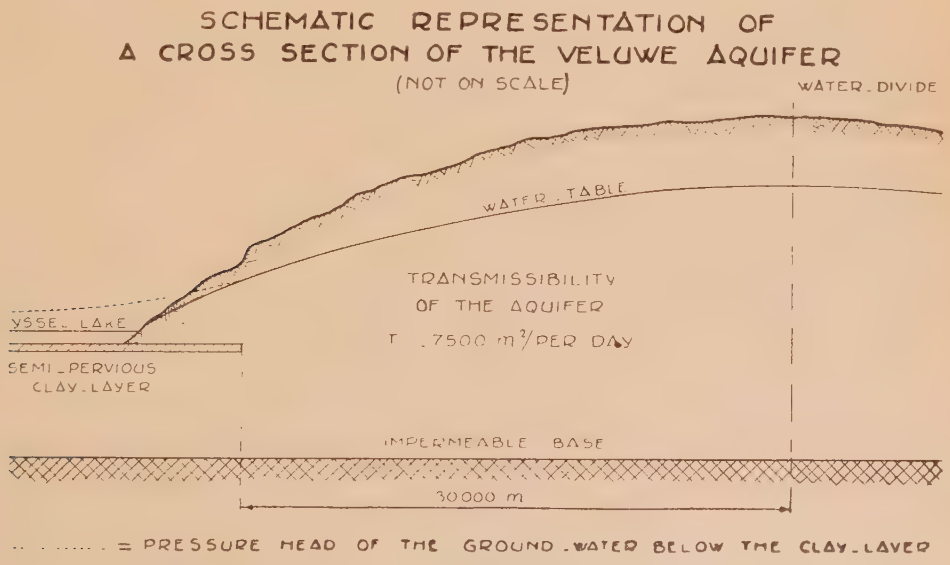

Fig. 2.

may best be approximated by a precipitation, the intensity of which varies as a sine with a period of a year and an amplitude of 2 mm rain per day.

The portion of the Veluwe which drains into the Yssel Lake covers an area of some 800 sq. km. From the coast this area extends inland over a distance of from 20 to 30 km.

The aquifer is not in open communication with the Yssel Lake : between them there is a semi-pervious cley-layer, offering a certain resistance to the discharge of gruond-water into the lake.A schematic representation of the situation is given in figure 2. From hydrologic investigations 0.16 days /m per running m of the Veluwe-coastline has been found to be the most probable value of the resistance of the semi-pervious clay-layer.

The problem of the variations in the discharge of the ground-water from the Veluwe into the Yssel Lake can be solved without difficulties only for the case where the flow of the ground-water is two-dimensional. The investigations, therefore, have been carried out only for the area of the Veluwe where the ground-water flow to the Yssel Lake is practically two-dimensional. The mean rate of discharge from this area has been computed ot 16 cubic meters per day per running m of coastal length of the Yssel Lake.

Since the Veluwe-aquifer is sufficiently homogeneous to consider its transmissibility as being constant, the problem has been reduced to the problem of the two-dimensional flow of ground-water in a homogeneous aquifer, which is continuously being replenished by rainfall, the intensity of which varies as a sine with a period of a year, the discharge of the ground-water from the aquifer into a lake taking place through a clay-layer.

IV. The discharge of ground-water from an area of infinite extent the intensity of the precipitation over this area varying as a sine.

The non-steady flow of ground-water in a homogeneous aquifer of constant transmissibility T, which is continuously being replenished by a variable precipitation N, is governed by the formula :

$$T \frac{\delta^2 z}{\partial x^2} + N = \beta \frac{\delta z}{\partial t} \qquad (1)$$

where z is the height of the water-table above the impermeable base of the aquifer at the time t and at the distance x from the zone of discharge; β the coefficient of porosity or the coefficient of storage of the soil under phreatic conditions.

The equation for the mean flow of the ground-water is

$$T \frac{\delta^2 z_m}{\delta x^2} + N_m = 0 \qquad (2)$$

in which z_m represents the mean height of the water-table at the distance x from the zone of discharge (the coast) and N_m is the mean value of the precipitation. Subtracting (1) and (2), we obtain the equation for the variations :

$$T \frac{\partial^2 z_v}{\partial x^2} + N_v = \beta \frac{\partial z_v}{\partial t} \qquad (3)$$

where $z_v = z - z_m$

$N_v = N - N_m$.

If we write

$$q = \frac{1}{\beta} \int N_v dt, \qquad (4)$$

in which q is a periodical function of the time t, we find after substitution of (4) in (3)

$$T \frac{\partial^2 z_v}{\partial x^2} = \beta \frac{\partial (z_v - q)}{\partial t} \qquad (5)$$

or, if we put

$$z_v - q = w \qquad (6)$$

$$\text{and } \frac{\beta}{T} = a, \qquad (7)$$

(1x) The figures refer to the references at the end of the paper.

we have

$$\frac{\partial^2 w}{\partial x^2} = a\,\frac{\partial w}{\partial t} \tag{8}$$

This method has been described in « Methoden der mathematischen Physik I» by R. Courant and D. Hilbert (1x), 2nd edition, p. 236 ...... « Eine homogene Differentialgleichung mit unhomogenen Randbedingungen ist äquivalent einer unhomogenen Differentialgleichung mit homogenen Randbedingungen. » (A homogeneous differential equation with non-homogeneous boundary conditions is equivalent to a non-homogeneous differential equation with homogeneous boundary conditions.)

In the simplified case of sinusoidally varying precipitation with a period of a year, the functions q and w are also sinusoidal with the same period and we may write

$$q = -A \sin \theta t \tag{9}$$

From (6) we find the boundary conditions for w :
At the discharge end of the aquifer, where $x = 0$, there is no fluctuation in the water level, so

$$z_v = 0 \text{ and } w = -q = A \sin \theta t \tag{10a}$$

At the water-divide of the aquifer where $x = L$, the ground-water gradient is zero, so

$$\frac{\delta w}{\partial x} = 0 \tag{10b}$$

The problem of the solution of w from (8), with the boundary conditions (10a) and (10b), corresponds entirely with the problem of the propagation of a tidal fluctuation w in ground-water. Both Forchheimer 2) and Steggewentz 3), regarding the effect of tidal oscillations in seas and estuaries on the ground-water levels in an adjacent aquifer of infinite extent, obtained the solution

$$w = Ae^{-x\sqrt{\frac{\theta a}{2}}} \sin\left(\theta\, t - x\sqrt{\frac{\theta a}{2}}\right) \tag{11}$$

For the case, where the aquifer is not infinite, the solution becomes

$$w = \frac{A}{\sqrt{1 + e^{-4L\sqrt{\frac{\theta a}{2}}} + 2e^{-2L\sqrt{\frac{\theta a}{2}}} \cos 2L\sqrt{\frac{\theta a}{2}}}}$$

$$\left[e^{-x\sqrt{\frac{\theta a}{2}}} \sin \left\{ \theta t - x\sqrt{\frac{\theta a}{2}} \right\} + e^{(x-2L)\sqrt{\frac{\theta a}{2}}} \sin \left\{ \theta t - (2L - x)\sqrt{\frac{\theta a}{2}} \right. \right. \tag{12}$$

where the boundary condition for $x = 0$ takes the form

$$w = \frac{A}{\sqrt{1 + e^{-4L\sqrt{\frac{\theta a}{2}}} + 2e^{-2L\sqrt{\frac{\theta a}{2}}} \cos 2L\sqrt{\frac{\theta a}{2}}}}$$

$$\left[\sin \theta t + e^{-2L\sqrt{\frac{\theta a}{2}}} \sin\left(\theta t - 2L\sqrt{\frac{\theta a}{2}}\right) \right] \tag{13}$$

In the case of the Veluwe -aquifer we may substitute

$$L = 30.000 \text{ m};$$

$$a = \frac{\beta}{T} = \frac{0,3}{7500} \text{ day}/\text{m}^2$$

Further, since the period of the fluctuations is a year, we have

$$\theta = \frac{2\pi}{360} \text{ day}^{-1}$$

Hence
$$e^{-2L\sqrt{\frac{\theta a}{2}}} = 2,5.10^{-16}.$$

This term is so small as to be negligible in (12) and (13). Then (12) is reduced to

(11), and, so far as the propagation of a « tide « with a period of a year is concerned, apparently the Veluwe-aquifer may be regarded as infinite. So formula (11) may be applied for solving the problem and finding w.

The varying amount of ground-water flowing through a cross-section of the aquifer of unit-width is

$$Q = T\frac{\partial z}{\partial x} = T\frac{\partial w}{\partial x} = -TA\sqrt{\frac{\theta a}{2}}\, e^{-x\sqrt{\frac{\theta a}{2}}}$$

$$\left\{\sin\left(\theta t - x\sqrt{\frac{\theta a}{2}}\right) + \cos\left(\theta t - x\sqrt{\frac{\theta a}{2}}\right)\right\} \qquad (14)$$

and, when $x = 0$

$$Q = -TA\sqrt{\frac{\theta a}{2}}\,(\sin\theta t + \cos\theta t) \qquad (15)$$

In the case where the aquifer at the point $x = 0$ is in communication with the open water only through a clay-layer with a resistance R, a loss of head occurs in this layer, which may be written

$$-TAR\sqrt{\frac{\theta a}{2}}\,(\sin\theta t + \cos\theta t) \qquad (16)$$

From (10a) and (16) we find the « tidal oscillation » of the open water

$$A\left\{\sin\theta t + TR\sqrt{\frac{\theta a}{2}}\,(\sin\theta t + \cos\theta t)\right\} =$$

$$A\sqrt{1 + T^2R^2\,\theta a + TR\sqrt{2\,\theta a}}$$

$$\sin\left\{\theta t + \text{arc cot}\left(1 + \frac{\sqrt{2}}{TR\sqrt{\theta a}}\right)\right\} \qquad (17)$$

If we introduce as a boundary condition for the open water

$$w = A^1 \sin\,\theta t \qquad (18)$$

the solution becomes

$$w = \frac{A^1}{\sqrt{1 + T^2R^2\theta a + TR\sqrt{2\,\theta a}}}\, e^{-x\sqrt{\frac{\theta a}{2}}}$$

$$\sin\left\{\theta t - \text{arc cot}\left(1 + \frac{\sqrt{2}}{TR\sqrt{\theta a}} - x\sqrt{\frac{\theta a}{2}}\right)\right\} \qquad (19)$$

which formula replaces (11).

It follows that the variation in the discharge of ground-water at the coast is

$$Q = \frac{-TA^1\sqrt{\theta a}}{\sqrt{1 + T^2R^2\,\theta a + TR\sqrt{2\,\theta a}}}$$

$$\sin\left\{\theta t + \frac{\pi}{4} - \text{auc cot}\left(1 + \frac{\sqrt{2}}{TR\sqrt{\theta a}}\right)\right\} \qquad (20)$$

Formula (19) may be transformed to an expression for z, the variation in the ground-water level, by means of (6).

V. Application of the derived formulas to the case of the Veluwe-discharge.

The formulas to be used in finding the variations in the Veluwe-discharge are :

for the fluctuations of the ground-water level caused by the precipitation : formula (4);

for the fluctuations caused by the imaginary tide : formules (18) and (19);

for the resulting fluctuation in the ground-water level : formula (6).;

and for the variation in the discharge at the coast : formula (20).
Further we may use the following data :

$$a = \frac{\beta}{T} = \frac{0,3}{7500} \text{ day /m}^2$$

R = 0,16 day /m per running m of the coast.
For x = 0 wa have $z_v = 0$ and from (6)

$$-q = w$$

On substituting w formula (18) becomes

$$q = -A^1 \sin \theta t = A^1 \sin (\theta t + \pi) \tag{21}$$

Hence, (4) may be written

$$A^I \sin (\theta t + \pi) = \frac{1}{\beta} \int N_v \, dt \qquad \text{or}$$

$$N_v = A^1 \theta \beta \sin \left(\theta t + \frac{3\pi}{2} \right) \tag{22}$$

It be assumed that the precipitation varies sinusoidally with a period of a year and an amplitude of 2 mm /day. We then obtain

$$| A^1 | = \frac{| N_v |}{\theta \beta} = 0,191 \text{ m.}$$

From (19) we now may find the « tidal » fluctuation w with its phase-angle for different values of the distance x. It should be noted that the distances are not measured from coast but from the edge of the clay-layer separating the Veluwe-aquifer from the Yssel Lake (fig. 2).

For reasons of simplicity, the distances have been so chosen, that the damping fector $D = e^{-x\sqrt{\frac{\theta a}{2}}}$ consecutively obtains the values 1, 4 /5, 3 /5, 2 /5, 1 /5, 1 /10 and 1 /20. The results of the calculation are given in the following table.

D	x	w	Phase-angle
1	0 m	0,103 m	— 22° 30′
4–5	380	0,083	— 35° 20′
3–5	865	0,062	— 51° 45′
2–5	1550	0,041	— 75°
1–5	2730	0,021	— 114° 55′
1–10	3900	0,010	— 154° 30′
1–20	5080	0,005	— 192° 30′

$$\text{Since } z_v = w + q \tag{6}$$

we may find the real fluctuations in the ground-water level through composing w and q. This may be done graphically by setting out in a vector-diagram the calculated values of w and composing them with the vector of q given by (21). The results obtained that way are given in the next table.

x	z_v	Phase-angle with respect to N_v
0 m	0,103 m	— 67° 30′
380	0,133	— 69°
865	0,160	— 72° 20′
1550	0,185	— 77° 40′
2730	0,200	— 84° 30′
3900	0,200	— 88° 30′
5080	0,195	— 90° 20′

It appears that the influence of the open water on the ground-water fluctuations caused by the precipitation decreases rapidly with an increasing distance x.

At a distance of some 5 km from the coast the influence is hardly noticeable; the difference between the calculated fluctuation z_v ($= 0{,}195$ m) and the undisturbed fluctuation q caused by the precipitation ($= 0{,}191$ m) is negligible.

Finally we find from (20) the variation in the discharge of ground-water at the coast

$$Q = 0{,}645 \text{ m}^3/\text{day per running m}'.$$

The phase-angle for Q may be calculated from

$$-\sin\left\{\theta t + \frac{4}{\pi} - \text{arc cot}\left(1 + \frac{\sqrt{2}}{TR\sqrt{\theta a}}\right)\right.$$

$$\text{or} \qquad -\sin\,(22^\circ\,30' + 180^\circ).$$

It follows from (22) that the difference in phase between Q and N_v is — 67° 30′, so Q appears to be in phase with the ground-water fluctuation z_v at the coast, as could be expected.

A phase difference of 67° 30′ corresponds with a length of time of $2\,^1/_4$ months. This leads to the conclusion that the effect of a dry period is perceptible in the discharge of ground-water at the coast only $2\,^1/_4$ months after the beginning of the arid period.

VI. Model investigations

The principle on which the model experiments were based may be explained as follows.

A viscous liquid passing laminarily between two parallel plates mounted together with an interspace of capillary dimensions, will encounter a linear resistance, according to the law of Poiseuille, Likewise ground-water passing through the soil, encounters a linear resistance according to Darcy's law. The conformity between these two laws enables us to imitate the motion of ground-water by means of the flow of a viscous liquid through an experimental model, consisting of two parallel plates put close together. Dachler (4) makes mention of a small apparatus constructed after this principle at the Technical University of Vienna. Further Uppal and Ahmad (5) have been using a model of this kind lately. A schematic front view of the model used in the laboratory of the State Institute for Water Supply is given in Fig. 3. The front and the back of the tank are transparent glass plates fixed together at an interspace of 2 mm. The length of the model is 2 m, the height 0,20 m. The liquid used is viscous paraffine oil which possesses the most suitable viscosity.

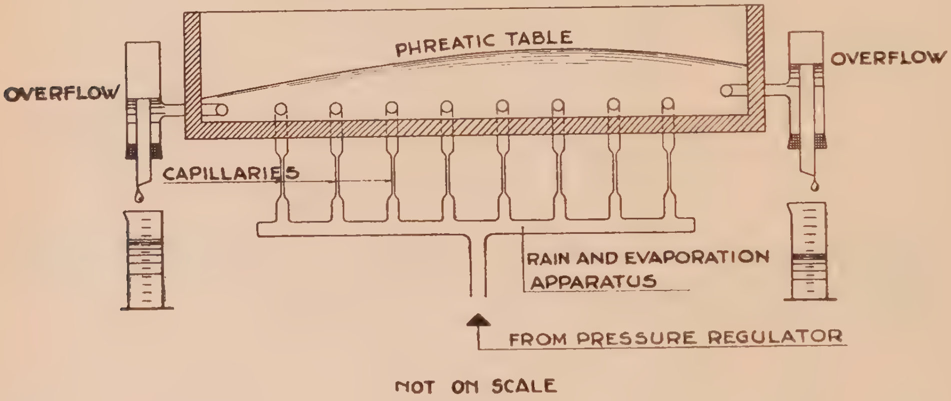

FIG. 3.

The flow of the paraffine oil through the tank is governed by the equation of Poisseuille

$$Q_m = \frac{1}{12}\, g\, \frac{b^3}{\nu}\, z_m \frac{\delta z_m}{\delta x_m} \qquad (23)$$

where

Q_m denotes the amount of oil moving through a cross section of the model;
g = the acceleration by gravity;
b = the distance between the parallel plates;
ν = the kinematic coefficient of viscosity of the oil;
z_m = the height of the oil surface above the bottom of the tank;
x_m = the horizontal co-ordinate.

The index m denotes that the factors, provided with is, refer to the model.

For the case where the variations in z_m are very small in respect tot the mean value of z_m, we may put

$$\frac{1}{12}\, g\, \frac{b^2}{\nu}\, z_m = T_m \tag{24}$$

where T_m may be defined as the transmissibility per unit-width of the model. Substituting in (23) gives

$$Q_m = b\, T_m \frac{\partial z_m}{\partial x_m} \tag{25}$$

Since (25) shows complete similarity with the general formula for the motion of ground-water,

$$Q = T \frac{\partial z}{\partial x},$$

it follows that we may use the model for investigation into ground-water problems with reservations for however some particular cases.

The model represents a cross-section of the Veluwe -aquifer, the paraffine oil in the model representing the body of ground-water above the impervious substratum. In contrast with the calculations given in the paragraphs IV and V, no use has been made in the model-experiment of the correspondance between the effect of sinusoidally varying precipitation and tidal propagation in ground-water but the actual rainfall has been imitated in the model. The incidence of rain may be imitated by a supply of paraffine, which is spread uniformly all along the full length of the model by means of a series of capillaries connected to the model. The amount of rain to be supplied may be regulated by means of a pressure regulator.

Overflow vessels, connected with each and of the model and made vertically adjustable, serve to regulate the level of the liquid at the ends, according to the boundary conditions of the ground-water flow.

A more complete description of the model has been given by Dietz 6).

The basic formula governing the non-steady motion of the ground-water in the aquifer is

$$T \frac{\partial^2 z}{\partial x^2} + N = \beta \frac{\partial z}{\partial t} \tag{1}$$

from which the equation for the model-scales may be derived:

$$\lambda_T \frac{\lambda_z}{\lambda^2_x} = \lambda_N = \lambda_\beta \frac{\lambda_z}{\lambda_t} \tag{28}$$

As followed from the calculations made in §V, the disturbing influence of the coast on the fluctuations of the ground-water caused by the precipitation decreases with the distance from the coast. At a distance of 5 km the influence has become negligible. So the investigations into the effect of the coast on the fluctuations of the ground-water may be restricted to the area within a distance of 5 km from the Yssel Lake. Therefore the model does not represent a cross-section of the whole Veluwe-aquifer, but only the coastal area unto a distance of 5 km from the coast. The length of the model being 2 m, we find

$$\lambda_x = 2500.$$

Since the porosity of the soil may be put approximately at 1/3, consequently $\lambda_\beta = 1/3$. Further λ_T follows from the ratio between T and T_m, T being 7500 m²/day, T_m being given by (24). When finally we choose $\lambda_z = 100$, the other scales $_N$ and λ_t may be found from (26).

At the end representing the coast, the model is in communication with an overflow vessel in which a constant level is being maintained, the level of the Yssel Lake. The resistance in the connection tube between model and overflow vessel corresponds with the resistance of the clay-layer between the open water of the Yssel Lake and the Veluwe-aquifer.

The other end of the model corresponds with a point situated inland at a distance of 5 km from the coast. As has already been mentioned here the fluctuation of the ground-water is practically undisturbed, the effect of the coast on the fluctuations at this distance being negligible. The second overflow vessel, therefore, which is in open communication with the model at this end, has to be moved up and down periodically with an amplitude of $2 \times 0{,}191$ m : λ_z or 3,82 mm, thus imitating the undisturbed groundwater fluctuations by the precipitation.

From (21) and (22) we see that a lag in phase of 90° or 3 months exists between the precipitation and the undisturbed ground-water fluctuation. Consequently the periodic motion given to the overflow vessel must lag 90° in phase with respect to the precipitation supplied to the model.

During the model-experiments the variations in the level of the paraffine oil in the tank were read from metal measuring tapes, fixed to the front plate of the model. Since the scale of the level variations, λ_z, is 100, a misreading of 0,1 mm corresponds with an error of 10 mm in reality.

The variations in the discharge of ground-water at the coast could be found by measuring the varying quantity of oil flowing out of the model. As this quantity was so small that the discharge took place in the form of drops, the variations in the discharge could be found by simply registrering every falling drop on a recording drum slowly turning with known angular velocity. Naturally the volume of a drop of paraffine oil had to be determined beforehand under different conditions of temperature and velocity of discharge.

The experiment lasted 3 × 9 min. 12 sec., representing a period of three years. From the readings and the other data obtained, the amplitudes and phase-angles of the level fluctuations and the discharge could be calculated. In the following table the results thus obtained are given together with the results found in chapter V by means of the tidal formulas.

	Ground-water fluctuation		Phase-angle with respect to the precipitation	
Distance from the edge of the clay-layer	found by the model experiment	calculated with tidal formulas	found by the model experiment	calculated from the tidal formulas
0 m	0,109 m	0,103 m	— 69° 10′	— 67° 30′
380	0,135	0,133	— 71° 5′	— 69°
865	0,156	0,160	— 70° 45′	— 72° 20′
1550	0,186	0,185	— 77° 40′	— 77° 40′
2730	0,199	0,200	— 82°	— 84° 30′
3900	0,195	0,200	— 86° 20′	— 88° 30′
	Variation in the ground-water discharge at the coast		Phaseangle with respect to the precipitation	
	0,658 m^3/day per m′	0,645 m^3/day per m′	— 66°	— 67° 30′

The deviation of the experimental level variations from the calculated values appears to be only a few percent. The phase-angles also agree very well. As a phase-angle of 1° corresponds with a period of 1 day or, on the time-scale of the model, $1\frac{1}{2}$ sec., a delay in the model-readings of only a few seconds will cause already a deviation from the correct phase-angle of some 2°. So all deviations can be explained fully from reading-errors.

It may be concluded, therefore, that both the tidal calculations and the model-experiment give results which are sufficiently accurate.

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2) VARIATIONS DU NIVEAU D'EAU ET DE LA PRESSION ASCENSIONNELLE DANS LES PUITS LEUR MESURE ET LEUR INTERPRETATION

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Les phénomènes constatés sur les forages de captage s'alimentant dans les nappes du crétacé inférieur du bassin parisien (albiennes principalement) illustrent d'une façon particulièrement brillante la relation étroite qui existe entre les variations du niveau de l'eau dans les puits s'alimentant dans une même nappe aquifère et celles de la pression à l'intérieur de cette dernière.

Ces nappes sont trop connues pour qu'il soit nécessaire d'en faire une description très détaillee; néanmoins, pour l'intelligence du présent rapport, nous rappelons brièvement, ci-après, leurs caractéristiques essentielles.

Parfois désignées sous le nom de « nappes de Gault » ou des « Sables Verts », elles sont incluses dans les assises sableuses, riches en glauconie, comprises entre les horizons argileux qui limitent, à la base et au sommet, les étages de l'Albien, de l'Aptien, voire du Barrémien ou du Néocomien, derniers termes du crétacé inférieur dans le bassin parisien (Fig. 1).

Ces différentes formations se présentent comme d'immenses cuvettes superposées, dont les parois, plissées par les phénomènes tectoniques qui ont affecté le bassin parisien, affleurent suivant une bande très perméable, presque continue entre la vallée de la Loire et les Ardennes, qui forme autour de la Capitale une auréole, bien caractérisée sur une carte géologique, qui traverse la Puisaye, la Champagne humide où sa largeur est maximum, le Barrois et l'Argonne.

Ces affleurements dont l'altitude moyenne est de 140 sont à un niveau toujours supérieur à la cote (+ 100), alors que le tréfonds des formations intéressées atteint, et dépasse peut-être en profondeur, la côte (— 800).

La puissance des sables verts est très variable. Toutefois, au fond de la cuvette albienne qui semble situé légèrement au nord-est de Paris, sur l'axe du synclinal de Saint-Denis, dans la région de Villemonble, vraisemblablement, les derniers forages exécutés ont révélé une épaisseur à peu près constante de 50 à 70 mètres, feuilletée par des intercalations argileuses qui subdivisent l'horizon aquifère en plusieurs nappes plus ou moins indépendantes.

Nous savons que celles-ci fournissent,dans la région parisienne, des eaux ayant une température de l'ordre de 30°, légèrement ferrugineuses et sulfureuses, d'un faible degré hydrométrique, et qui se sont montrées jusqu'à présent irréprochables au point de vue bactériologique.

L'alimentation de ces nappes a donné lieu à de nombreuses controverses. Cependant, on est à peu près d'accord pour admettre qu'elle est assurée par l'infiltration des eaux atmosphériques dans les affleurements du crétacé inférieur, sur les parties orientales du bassin parisien (sables de la Puisaye, de la Champagne humide et du Barrois). Cette opinion très raisonnable est également la plus vraisemblable. Elle paraît, d'autre part, avoir été confirmée par les expériences effectuées par Belgrand sur les eaux du puits de Grenelle, premier puits aménagé en vue de l'exploitation de ces nappes, expériences qui montrèrent la corrélation qui existe entre le degré hydrotimétrique des eaux fournies par les puits, c'est-à-dire leur teneur en sels alcalino-terreux, et les avalaisons aux affleurements.

Les eaux des nappes albienne et aptienne sont artésiennes, c'est-à-dire qu'elles remontent naturellement dans les forages construits en vue de leur captage. Celles qui vinrent au jour par les premiers ouvrages, auxquels leur exploitation donna lieu, étaient même jaillissantes.

Tous les ouvrages construits jusqu'à ce jour ont confirmé ce caractère essentiel qui s'explique, d'ailleurs, lorsque l'on considére les conditions de gisement des eaux intéressées. Ces dernières sont, en effet, captives entre deux assises, c'est-à-dire entre un « toit » et un « mur » imperméables. La partie inférieure de l'énorme cuvette qui

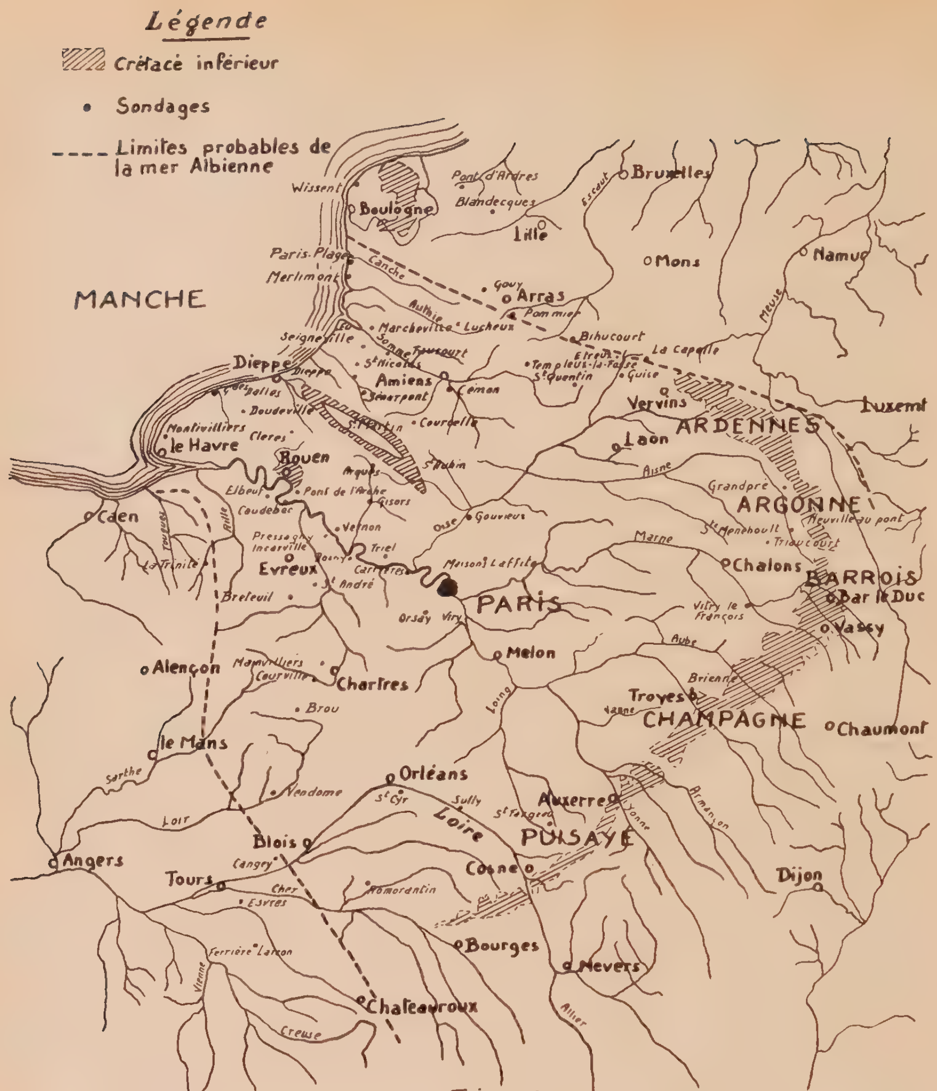

Fig. 1

existe sous la partie centrale du bassin parisien s'abaisse, comme nous l'avons vu ci-dessus, jusqu'au voisinage de la cote (— 800), alors que les bords se relèvent, par endroits, à (+ 200). De ce fait, les eaux incluses dans les horizons aquifères se trouvent sous une pression variable mais qui peut être considérable et atteindre une centaine de kilogs par centimètre carré pour celles situées au fond de la fosse.

Un ponction pratiquée au sein de la nappe entraîne dès lors la remontée immédiate des eaux incluses, et provoque, en suivant la description de Gérard (1) : « la puissante colonne d'eau qui jaillit spontanément au dessus de l'orifice, remontant » avec elle quantité de sable et de détritus provenant des terrains traversés, et qui » retomba en déluge sur les ouvriers... » ... lorsque, le 26 février 1841, vers 2 heures de l'après-midi, la sonde du puits de Grenelle perça définitivement les argiles de Gault et s'enfonça de plusieurs mètres dans les sables verts, portant la première atteinte à des nappes encore vierges et libérant des eaux enfouies au sein de la terre depuis des millénaires sans doute.

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(1) E. Gerard, *Paris souterrain*, page 269.

Exploitation de la nappe albienne

On vient de voir que l'exploitation de cette nappe commença en 1841 par la création du puits artésien de Grenelle. La profondeur totale atteinte à cette occasion était de 548 m. Le toit de sables verts fut rencontré à la profondeur de 527 m, l'altitude au sol étant de (+ 37 m). Le débit obtenu fut estimé à plus de 3.000 m^3 par 24 heures. Au moment où la sonde du foreur libéra la nappe, les eaux jaillirent au dessus du sol, à une certaine hauteur malheureusement non indiquée.

L'exploitation de la nappe fut continuée par le forage du puits dit de Passy (en 1855), dont le débit au sol (altitude + 53,15) fut d'abord estimé à 25.000 m^3 par 24 heures et s'abaissa progressivement jusqu'au chiffre de 6.000 m^3 par 24 heures (à l'altitude + 77).

Par la suite, fut construit le puits de la Butte-aux-Cailles, en 1863.

Puis se succédèrent un grand nombre de puits situés aussi bien dans la région parisienne que dans les départements du bassin parisien. Le plus grand nombre d'entre eux fut construit soit à Paris même, soit dans sa banlieue immédiate. Mais il y en eut aussi dans les départements de la Seine Inférieure, de Seine et Oise, de l'Indre et Loire, de l'Eure, du Pas de-Calais, etc... etc...

Le nombre total de forages ayant atteint la nappe albienne dépasse vraisemblablement une centaine, peut-être même un nombre supérieur.

Leurs caractéristiques sont variées : leur profondeur, en particulier, dépend de leur situation par rapport au centre et aux bords de la cuvette parisienne.

Le débit total extrait par ces ouvrages n'a jamais été connu avec précision. Certains l'ont estimé à 200.000 mètres cubes par jour, en 1934. Ce chiffre nous paraît exagéré et le débit global n'a sans doute jamais dépassé à un même moment, s'il l'a atteint, le chiffre de 100.000 mètres cubes par jour.

Le nombre de forages encore en service est faible et ne doit pas dépasser une dizaine. La plupart d'entre eux ne présentent plus guère d'intérêt, aux yeux de leurs propriétaires, du fait qu'ils ont perdu la qualité qu'ils estimaient primordiale : celle d'être jaillissants.

La disparition, toute naturelle d'ailleurs, de la propriété en question a été à l'origine des controverses scientifiques ou autres auxquelles ont donné lieu les nappes intéressées, et que nous allons examiner ci-après.

Phénomènes observés sur les forages du crétacé inférieur

Nous ne pouvons ,dans le présent rapport, nous livrer à une étude historique de la construction des forages de l'Albien, malgré tout l'intérêt qu'elle présente.

Cette étude montrerait, en effet, un accroissement rapide du nombre de ces ouvrages, au début de leur vogue, suivi d'un ralentissement à peu près total de leur construction, puis une reprise très active dans certaines régions (la région parisienne en particulier) lorsque furent introduites en France, vers 1930, les méthodes d'exécution américaines connues sous le nom de « forage au Rotary ».

La période initiale de développement intensif de ces forages eut pour origine les résultats spectaculaires obtenus dès la création des premiers ouvrages, résultats qui sont relatés dans les chroniques de l'époque. La période de ralentissement fut consécutive aux déboires auxquels donna lieu l'exploitation intensive et un peu désordonnée de ces même souvrages. Les usagers s'aperçurent avec un certain étonnement que les cotes de jaillissement constatées tout d'abord, ainsi que les débits très importants qui avaient été obtenus à l'origine, loin de se maintenir, diminuaient rapidement, au fur et à mesure de la création de nouveaux ouvrages ; bien plus, le forage d'un puits nouveau présentait le plus souvent une incidence directe, soit sur le débit, soit sur la hauteur de jaillissement des ouvrages voisins, soit sur ces deux grandeurs simultanément.

Ces phénomènes furent généralement attribués à l'épuisement de la nappe et les préoccupations qu'elles provoquèrent furent sans doute à l'origine du décret-loi du 8 août 1935 par lequel les pouvoirs publics jugèrent indispensables d'apporter des limitations à la création de nouveaux forages profonds. Cette législation, encore à ses débuts d'ailleurs, doit permettre de réserver les eaux souterraines profondes à la satisfaction des besoins collectifs les plus indispensables. Elle est dominée par la notion « d'épuisement » dont plusieurs géologues s'étaient faits les champions.

En nous penchant dès lors sur ce problème si important, nous avons été frappés par plusieurs faits, dont deux en particulier nous ont amené à émettre des doutes sur la thèse de l'épuisement telle qu'elle était présentée et prônée à l'époque. C'est ainsi que l'étude de la documentation, qu'il nous a été possible de réunir, mit en évidence le fait que les phénomènes de diminution de débit constatés sur certains ouvrages, au lieu d'avoir une allure continue en fonction du temps, procédaient au contraire par à-coups successifs qui coïncidaient à quelques heures près avec le forage de nouveaux puits dans leur voisinage.

D'autre part, si les réserves de la nappe avaient été directement en cause, tous les ouvrages implantés dans cette dernière auraient dû être intéressés par l'épuisement de celle-ci. De plus, le phénomène aurait dû s'accélérer au fur et a mesure que croissait le nombre d'ouvrages en exploitation. Or, bien au contraire, les réactions provoquées par la construction d'un nouvel ouurage étaient, en général, limitées aux forages de captage les plus proches. Par ailleurs, malgré l'augmentation du nombre d'ouvrages en exploitation, il était possible d'obtenir des derniers puits créés des débits du même ordre de grandeur, voire plus élevés, que des tout premiers puits, à la seule condition d'abaisser suffisamment le plan d'eau de pompage.

Car, durant toute cette période, si le caractère artésien des eaux incluses dans l'albien avait persisté, il n'en apparaissait pas moins que le niveau d'établissement s'était abaissé sensiblement, à telle enseigne que, quelques années déjà avant la deuxième guerre mondiale, il n'était guère possible d'obtenir à Paris même des ouvrages jaillissants proprement dits.

Pour corroborer ces indications, il n'est que de rappeler que :

1) Le débit du puits de Grenelle, qui était de 37 litres par seconde, au moment du jaillissement, s'abaissa à 13 litres par seconde lorsque l'on remonta le tubage de ce puits de 36 m 40 pour obtenir des eaux à la même pression, au-dessus du sol, que celle du réservoir de Panthéon ;

2) Dans une note présentée à l'académie des Sciences, en 1861, par J.B. Dumas, il est dit que :

» le débit observé au puits de Grenelle est resté, comme précédemment, de 900 m^3
» par 24 heures jusqu'au 25 septembre 1861 à midi ; mais, le même jour à minuit,
» il est tombé à 806 m^3 ; le 26 à 6 heures du soir, à 677 m^3,... etc... » ;

si l'on veut bien se rappeler que les eaux du puits artésien de Passy, situé à quelques kilomètres à peine à vol d'oiseau de celui de Grenelle, virent le jour le 24 septembre 1861, la relation de cause à effet semble évidente, les quelques heures écoulées entre les deux pnénomènes susvisés représentant le temps strictement nécessaire à la nappe albienne pour retrouver un nouvel état d'équilibre ;

3) Une communication à l'Académie des Sciences, de M. Huet, présentée par M. Daubret, au mois de juillet 1888, indique de même qu'« après le percement du puits de Passy, le débit du puits de Grenelle, qui était de 900 m^3 par 24 heures, est tombé à 650 m^3 ».

Il serait possible de multiplier les citations du même genre,mais elles se trouvent toutes implicetement contenues dans l'opinion émise par Belgrand, Inspecteur Général des Ponts et Chaussées, Directeur des Eaux et Egouts de Paris, Membre de l'Institut, quand il écrivit dans « Les Eaux Nouvelles » que :

» le forage d'un nouveau puits artésien a pour effet de réduire considérablement non
» seulement le débit, mais encore la hauteur du point hydrostatique ou l'altitude
» à laquelle on peut faire monter l'eau des puits. »

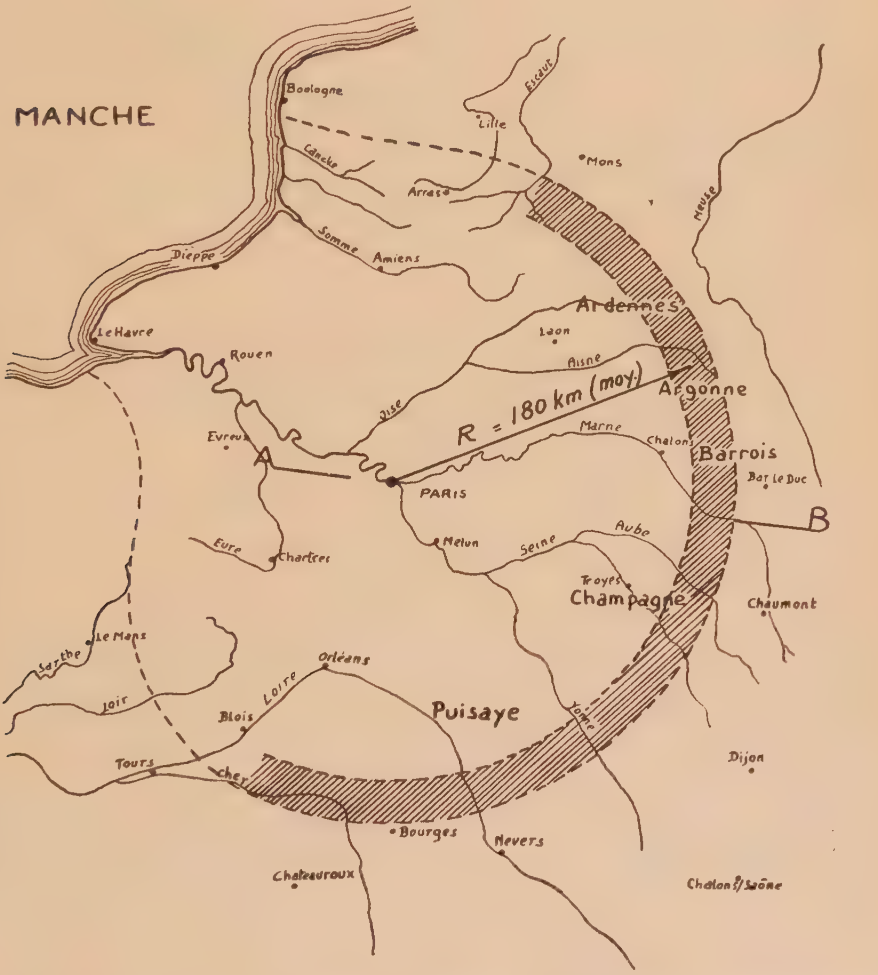
Des faits plus récents ont corroboré cette opinion. C'est ainsi que la courbe des débits du puits artésein de la Raffinerie Say, construit en 1869, accuse un abaissement brutal assez important vers 1903–1904, époque qui vit la venue au jour des eaux issues du puits de la Butte-aux-Cailles, commencé le 14 juillet 1863.

La concomitance ainsi bien établie des phénomènes observés à de nombreuses reprises contredit en l'espèce la théorie de l'épuisement et oblige à rechercher une autre interprétation de ces phénomènes. Or ceux-ci s'expliquent aisément si l'on veut bien tenir compte du fait que l'écoulement des eaux souterraines, sur tout le parcours de leur long trajet, depuis les affleurements du crétacé inférieur jusqu'au fond de la cuvette parisienne, ne peut s'effectuer sans donner lieu à des pertes de charge, pertes de charge qui doivent augmenter avec le débit demandé à la nappe et avoir pour conséquence l'abaissement du niveau piézométrique aux points de puisage.

Si, eu égard aux nombreux facteurs à considérer, et à la complexité de la morphologie d'une nappe aussi étendue que celle qui fait l'objet de la présente étude, il est difficile de donner une représentation quantitative des phénomènes, du moins peut-on, par une schématisation préalable, l'exprimer qualitativement.

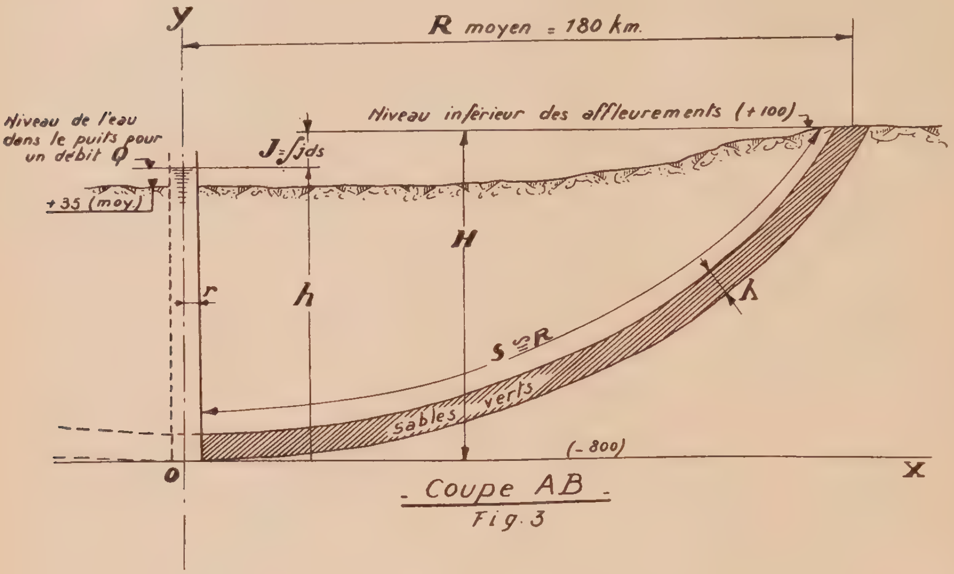
Dans ce but, considérons essentiellement les parties de la nappe situées, au point de vue de l'écoulement souterrain des eaux, en amont de Paris, dans les directions Est et Est-Sud-Est, qui sont d'ailleurs les plus intéressantes quant au comportement hydrodynamique des puits parisiens. Nous pouvons considérer dans de telles limites la nappe albienne comme formant une vaste dépression en forme de demi-calotte sphérique dont Paris serait le centre, et dont les bords ne seraient autres que les affleurements alimentaires se développant depuis Bourges et Romorantin, au Sud, jusqu'à Vervins et La Capelle, au Nord, suivant un vaste demi-cercle de 180 kilomètres de rayon (Fig. 2). Supposons de plus, par mesure de simplification, que cette nappe a une épaisseur uniforme λ, et représentons par H la hauteur moyenne des affleurements, par rapport à l'axe horozintal OX passant par la partie la plus profonde de la cuvette au droit de Paris.

Enfin, pour alléger schématiquement l'étude hydrodynamique de l'écoulement, et sans que cela puisse nuire en quoi que ce soit à la rigueur du raisonnement, nous admettrons que tous les puits construits dans la région parisienne immédiate se



Tracé schématique des affleurements du crétacé inférieur
Fig. 2

trouvent concentrés au centre même de la capitale, en un puits unique, ayant un diamètre $2r$ tel que la section correspondante soit proportionnelle au débit demandé au puits en question (Fig. 3).


Y
R moyen = 180 km.
Niveau de l'eau dans le puits pour un débit Q
+35 (moy.)
Niveau inférieur des affleurements (+100)
J=∫jds
H
r
h
h
s≅R
sables verts
(-800)
O
X

- Coupe AB -
Fig. 3

Si u est la vitesse moyenne de circulation aux affleurements correspondant au débit Q extrait de ce puits hypothétique,

U la vitesse de circulation à l'intérieur du tubage du puits,

P_a la pression atmosphérique,

ω le poids spécifique de l'eau extraite de la nappe,

$h<H$ la hauteur mesurée à partir de l'axe OX à laquelle cette eau est extraite du puits,

il est possible d'écrire avec une grande approximation :

$$\frac{u^2}{2g} + \frac{P_a}{\omega} + H = \frac{U^2}{2g} + \frac{P_a}{\omega} + h + \int_o^s jds \tag{1}$$

expression dans laquelle $\int_o^s jds$ représente la somme des pertes de charge sur le parcours du flux liquide depuis les affleurements jusqu'au puits central d'exhaure.

Or ,comme nous le savons déjà :
H est compris entre 800 et 1000 mètres.
h quoique plus petit est d'un ordre de grandeur comparable,
U vitesse ascensionnelle dans le tubage, est compris entre 1 et 2 mètres au maximum,
Quant à u, vitesse aux affleurements, il est au maximum de quelques centimètres par seconde.

Dans ces conditions, tous les termes de l'équation (1) représentant de l'énergie cinétique, sont négligeables devant H et h, de sorte que l'on peut écrire finalement :

$$\begin{aligned} H &= h + \int_o^s jds \\ &= h + J \\ \text{soit } J &= H - h \end{aligned}$$

J comprend non seulement les pertses de charge internes mais également celles qui se produisent dans le tubage de l'ouvrage captant, ainsi que les pertes de charge singulières qui sont dues, par exemple, à la mise en vitesse u du flux aquifère aux affleurements, au encore qui peuvent se produire ou moment où les eaux captées pénètrent à la base du puits. En tout état de cause, l'ensemble de celles-ci, dont la somme est de l'ordre de quelques centimètres au plus, apparaît comme négligeable devant les pertes de charge internes proprement dites qui, comme nous le verrons ci-après, peuvent atteindre des valeurs considérables. Nous nous bornerons dès lors à considérer ces dernières.

On admet très généralement (1) que le mouvement d'écoulement du flux liquide à l'intérieur de la nappe obéit au régime laminaire. Les pertes de charge sont donc liées à la vitesse par la relation classique de Darcy : $J = \alpha u$.

Si Q est le débit extrait du puits central et si l'on désigne comme d'habitude par μ le coefficient de vide au plein du milieu poreux dans lequel la nappe est incluse, on obtient au droit des affleurements la relation :

$$u = \frac{Q}{\mu \pi R \lambda}$$

De même dans une section circulaire concentrique à celle des affleurements et déterminée par sa distance X du centre O de la nappe, on peut écrire :

$$u_x = \frac{Q}{\mu \pi x \lambda}$$

vitesse a laquelle correspond une perte de charge unitaire ayant pour expression :

$$j_x = \alpha u_x = \frac{\alpha Q}{\mu \pi x \lambda} = \frac{Q}{\epsilon \pi x \lambda}$$

compte tenu de ce que $\frac{\mu}{\alpha} = \epsilon$ n'est autre que le coefficient caractéristique du terrain connu sous le nom de module de Thiem.

Pour l'ensemble de la nappe, nous pouvons écrire en conséquence :

$$J = \int_r^R \frac{\alpha}{\mu} \frac{Q\, dr}{\pi x \lambda} = \frac{Q}{\epsilon \pi \mu} \operatorname{Log}_e \frac{R}{r}$$

et si, compte tenu de ce que nous avons dit précédemment sur le puits équivalent, nous posons :

$$\pi r^2 = k\ Q$$

k étant une constante déterminée par la vitesse uniforme que l'on se propose de maintenir dans le tubage du puits central, ou encore :

$$r = \left(\frac{k}{\pi} Q\right)^{1/2} = K\, Q^{1/2}$$

il vient :

$$J = \frac{}{\epsilon \pi \lambda} \operatorname{Log}_e \frac{R}{K\, Q^{1/2}}$$

$$J = \frac{Q}{\epsilon \pi \lambda} \left[\operatorname{Log}_e R - \operatorname{Log}_e KQ^{1/2}\right]$$

expression qui montre que la perte de charge due à l'écoulement des eaux albiennes dans leur gisement, sans être rigoureusement proportionnelle au débit Q demandé à la nappe, augmente rapidement avec ce débit, l'influence pondératrice de la partie négative du terme entre crochets ($\operatorname{Log}_e K\, Q^{1/2}$) étant excessivement faible.

Il en découle que la hauteur piézométrique h à laquelle les eaux peuvent être captées au centre de la cuvette parisienne, soit :

$$h = H - \frac{Q}{\epsilon \pi \lambda} \left[\operatorname{Log}_e R - \operatorname{Log}_e K\, Q^{1/2}\right]$$

doit diminuer au fur et à mesure que le débit demandé à la nappe augmente, conclusion logique, en accord avec les lois de l'hydrodynamique et qui explique sans laisser la moindre obscurité tous les phénomènes constatés sur les différents puits ou forages s'alimentant dans les nappes du crétacé inférieur du Bassin parisien.

Elle suffit, en particulier, pour mettre hors de cause, à peu près complètement, la notion d'épuisement de la nappe albienne et par conséquent pour rendre sujets à caution les espoirs que l'on pourrait fonder sur une solution de réalimentation de la nappe.

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(1) Voir notament la justification de cette hypothèse dans : *Une vérification à grande echelle de la loi de Darcy*, par A. Vibert — Génie Civil du 22 juillet 1939; page 84.

Elle traduit à l'évidence le fait (1) qu'il existe une relation mécanique certaine et étroite entre les variation du niveau, du débit et de la pression en un point quelconque d'une nappe captive analogue à celle que nous nous sommes proposé d'étudier.

A titre d'approximation et pour tenter d'apprécier quantitativement les données du problème, on peut prendre pour les différents paramètres, qui entrent dans la formule à laquelle nous avons abouti, des valeurs moyennes dont l'ordre de grandeur soit celui des caractéristiques propres à la nappe étudiée.

C'est ainsi qu'en prenant le mètre, le mètre par seconde et le mètre cube pour unités respectives des longueurs, des vitesses et des débits, on peut poser :

$R = 180$ km $= 180.000$ mètres,
$\lambda = 32$ mètres,
$K = 1$ valeur correspondant à une vitesse ascensionnelle constante de 1 mètre dans le tubage du puits central, quel que soit le débit extrait de la nappe,
$\epsilon = 0,0005$ environ, ordre de grandeur valable (2) pour un milieu poreux composé par des sables fins comparables à ceux qui constituent les assises albiennes.

Il vient dès lors, avec $H = 100 + 800 = 900$:

$$h = 900 - \frac{Q}{0,0005 \times 3,14 \times 32} [\text{Log}_e\, 180.000 - \text{Log}_e\, Q^{1/2}]$$

et, comme $0,0005 \times 3,14 \times 32 \cong 0,05$,

$$h = 940 - 20\, Q\, [\text{Log}_e\, 180.000 - \text{Log}_e\, Q^{1/2}]$$

Cette fonction est représentée par la figure 4 qui ne prétend sans doute pas à une précision quantitative, mais qui donne, malgré tout, une image assez fidèle de ce qui s'est passé à Paris, si l'on considère qu'en 1935 le niveau pièzométrique de la nappe albienne avait baissé de 80 à 100 mètres depuis le début de son exploitation, alors que le débit total journalier extrait dans la région parisienne avait passé de quelques milliers à 60.000 mètres cubes par 24 heures en chiffres ronds.

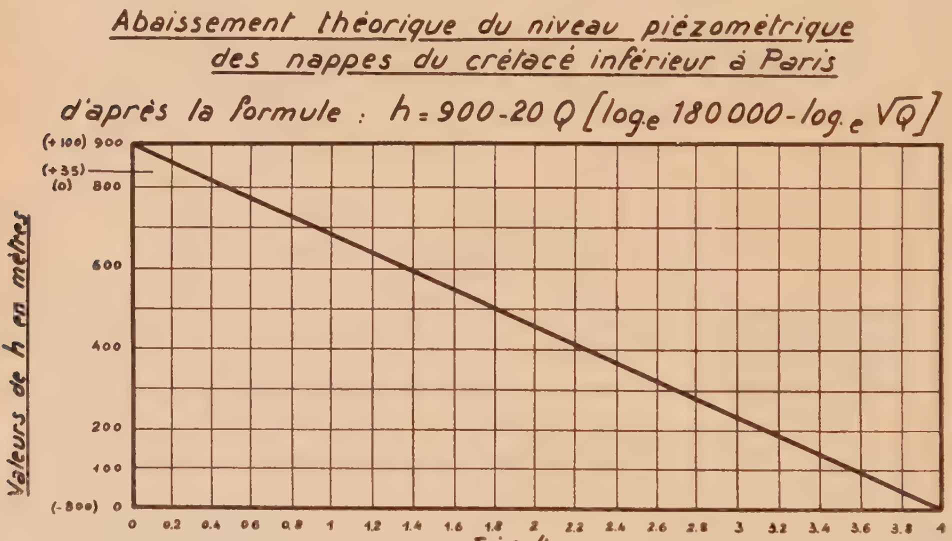

Fig. 4

Débits Q en m³/s

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(1) Ce qui était l'objet même de la deuxième question posée au Congrès de l'A.I.H.S.

(2) D'après Imbeaux, *Essai d'hydrologie*, page 62, ϵ peut même descendre bien au-dessous de 0,0003 jusqu'à atteindre 0,0005 pour des sables limoneux.

L'interprétation plus étendue du même graphique montrerait que, pour un débit de l'ordre de 4 mètres cubes /seconde, soit 350.000 mètres cubes par jour, l'artésianisme de ces nappes aurait complètement disparu et que les eaux devraient être puisées au fond même de cette cuvette, ces chiffres n'ayant toutefois de valeur que dans la mesure où les hypothèses « moyennes » qui ont été faites plus haut se traduiraient de la réalité.

Mais, si la grandeur de ces phénomènes reste discutable, nous croyons avoir montré que leur signification ne l'est pas et peut être précisée par des considérations exclusivement hydrodynamiques.

3) ETUDE DE LA SURFACE LIBRE DES NAPPES PAR LES CARTES DES COURBES ISOPIEZOMETRIQUES ECOULEMENT, ALIMENTATION ET SEUILS HYDRAULIQUES

par

G. CASTANY, chef du service géologique.

Cette note a pour but de présenter deux études actuellement importantes pour l'hydraulique souterraine de la Tunisie :

I. — Les caractéristiques des nappes quaternaires de Kairouan et Grombalia;

II. — Le problème soulevé par une série d'accidents tectoniques analogues qui, bien que peu marqués dans la topographie, occasionnent des chutes de niveau piézométrique et partagent la Tunisie Centrale en une série de cuvettes hydrauliques situées à des niveaux différents.

Elles nous ont conduit à dresser avec précision et à étudier en détail les cartes des courbes isopiézométriques.

L'étude hydrostatique d'une *nappe libre* peut se traduire dans un nombre d'observations suffisant, par une carte en courbes qui représente la forme de la surface supérieure de la nappe. Nous désignerons avec M. M. GOSSELIN par le terme de *courbes isopiézométriques* les courbes de niveau de son toit qui, dans ce cas, correspond à la *surface libre.*

Cette représentation permettant de connaître la forme de cette surface donne lieu à des observations précises sur l'alimentation et l'écoulement de l'horizon aquifère.

I. — *Caractéristiques des nappes quaternaires des régions de Kairouan et Grombalia* (fig. 1).

Nous étudierons successivement :
A) — Cuvette de Djébibina — Kairouan — Sidi el Hani;
B) — Golfe d'Enfidaville — Lac Kelbia;
C) — Plaine de Grombalia.

A.) — *Cuvette de Djébibina — Kairouan — Sidi el Hani.*

La cuvette de Djébibina, Kairouan, Sidi el Hani est située en bordure des chaînes plissées de l'Atlas Tunisien. Elle s'étend, du Nord au Sud, sur une centaine de kilomètres. Sa partie la plus large, d'Est en Ouest, dépasse 80 kilomètres. Elle est occupée par des formations quaternaires récentes, à couches plus ou moins sableuses donnant lieu à un ensemble de nappes aquifères de niveaux statiques légèrement différents.

Les courbes isopiézométriques tracées sur la carte ci-jointe (fig.1) indiquent le niveau de la nappe phréatique. Ces courbes épousent en gros la forme des courbes de niveau topographique et font ressortir les zônes de drainage du système hydraulique, qui sont les suivantes :

Bas-fonds d'El Mesbasta, Ouled en Nsar, Oued el Ataf. L'eau affleure au niveau du sol dans ces points bas. Elle est reprise par une évaporation intense.

La partie qui échappe à l'évaporation est drainée par l'Oued el Ataf et son underflow vers la Sebkha Kelbia.

Sebkhas de Sidi el Hani et de Chérita où l'eau est également reprise par l'évaporation.

On peut distinguer trois secteurs dans la cuvette drainée par ces exutoires :

Le secteur Nord, au Nord des bas fonds d'El Mesbasta dont les eaux s'écoulent du Nord au Sud.

Le secteur central dont les eaux s'écoulent partie au Nord vers el Masbasta, partie au Sud vers les sebkhas.

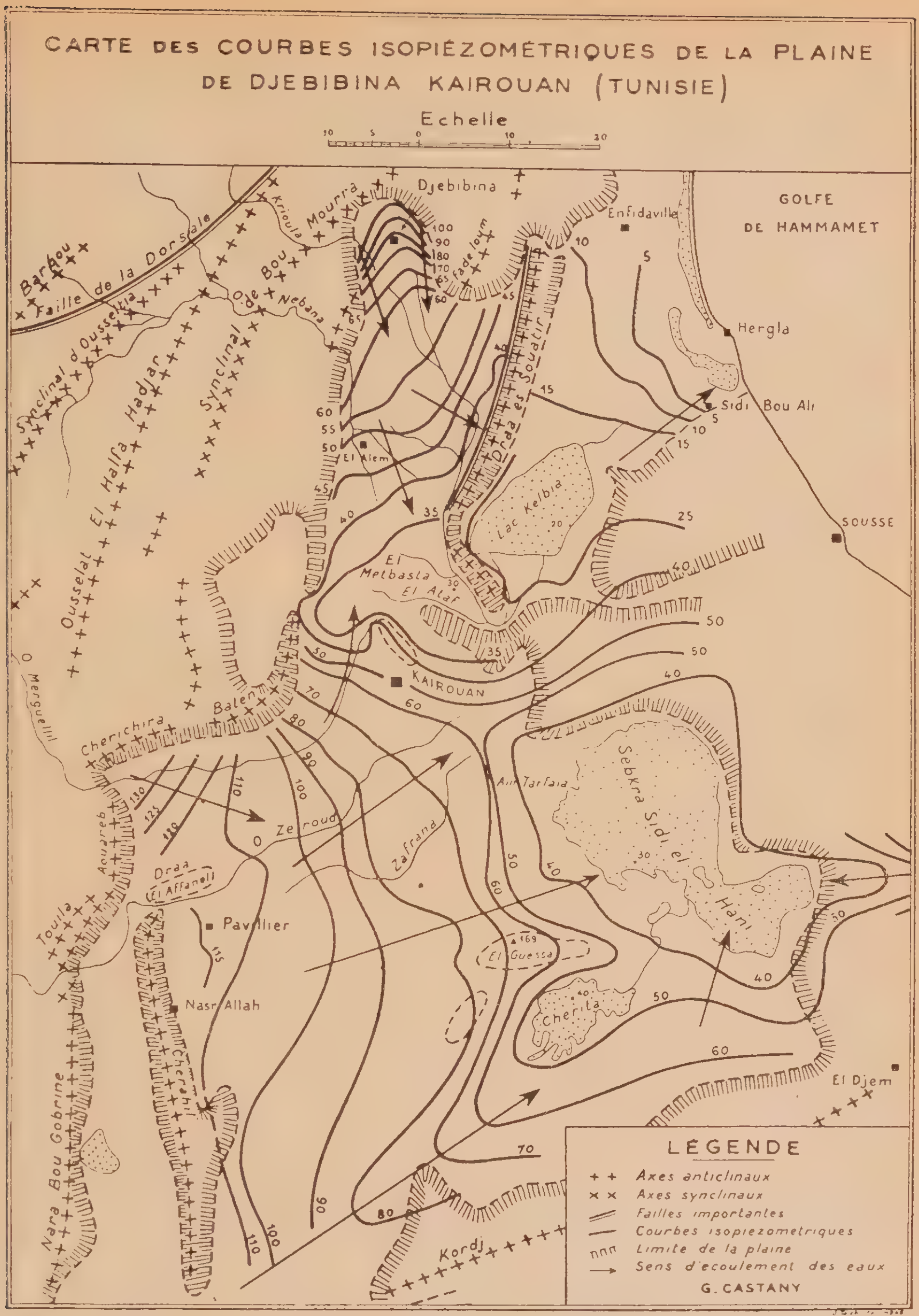

CARTE DES COURBES ISOPIÉZOMÉTRIQUES DE LA PLAINE DE DJEBIBINA KAIROUAN (TUNISIE)
Echelle
10 5 0 10 20
GOLFE DE HAMMAMET
Djebibina
Enfidaville
Hergla
Sidi Bou Ali
SOUSSE
KAIROUAN
El Alem
El Metbasta
El Alaf
Lac Kelbia
Sebkra Sidi el Hani
Ain Tarfaia
Zeroud
Zafrana
El Guessa
Cherita
Pavillier
Nasr Allah
El Djem
Kordj
Barbou
Faille de la Dorsale
Synclinal d'Ousselat
Ousselat
El Halfa Hadjar
Synclinal de Bou Nebana
Krioula
Mourra
Fadeloum
Draa es Souatir
Merguellil
Cherichira
Balen
Aousreb
Draa El Affanett
Touila
Nara Bou Gobrine
Cherahil
LÉGENDE
+ + Axes anticlinaux
× × Axes synclinaux
Failles importantes
Courbes isopiézométriques
Limite de la plaine
Sens d'écoulement des eaux
G. CASTANY

FIG. 1

Le secteur Sud, drainé vers les sebkhas.

1) — *Secteur Nord.* L'examen des courbes isopiézométriques (fig. 1) montre que la surface libre n'est pas régulière et qu'elle présente de nombreuses et importantes anomalies.

Le point haut de la nappe est au Nord à la cote 100. Vers le Sud, la surface piézométrique s'équilibre avec la plaine d'El Metbasta à la cote 25.

De là les eaux qui n'ont pas été évacuées soit souterrainement, soit par l'Oued el Ataf, vers la cuvette d'Enfidaville, sont absorbées par évaporation.

Au Nord, dans la plaine de Djébibina les courbes isopiézométriques très rapprochées et régulières présentent une forte concavité tournée vers les cotes décroissantes et en direction du Sud qui représente le sens du courant général. *La pente hydraulique est très forte*, égale à 4 pour 1000.

Les nombreux sondages forés dans cette région ont mis en évidence la présence d'un Quaternaire relativement perméable à lits de galets et de sables, et d'autant plus grossiers que l'on se rapproche des bordures. « Le gradient hydraulique » et la perméabilité indiquent un *fort débit* et par suite *une alimentation importante*. La courbe 60 montre une venue d'eau latérale occidentale très forte (courbe tangente aux affleurements de bordure). Elle est due aux grès oligocènes, miocène inférieur, et au Quaternaire du synclinal de Bou Mourra que viennent renforcer les crues et les underflows des Oueds issus de la Dorsale Tunisienne. Cette hypothèse est d'ailleurs confirmée par les études chimiques.

L'écoulement de la nappe s'effectue surtout par les bordures, phénomène qui, s'ajoutant à l'accroissement de l'alimentation dans ces zones, explique *l'allure en croissant du faisceau de courbes.*

Au Sud : On peut distinguer :

— un bassin faisant suite à la zone précédente où le sens d'écoulement des eaux s'incurve vers l'Est pour aboutir à un niveau de drainage longeant la colline du Draa es Souatir, et correspondant aux mouillères et au lit de l'Oued Bassaas. Les courbes sont plus espacées et la pente hydraulique moins rapide, 1 pour 1000. La concavité est tournée vers le Nord et l'axe de cette déformation dirigé N.O.-S.E. prolonge sensiblement celui des courbes septentrionales qui, nous l'avons noté, se caractérisaient par une pente rapide. Les sondages et l'examen des coupes dans les puits montrent que le terrain, du moins en surface, est *nettement moins perméable qu'au Nord.* Ce colmatage s'accroit encore vers le Sud. En même temps la salure des eaux phréatiques augmente et passe de 1 gr. par litre à 3 gr. environ. *L'écoulement de la nappe est donc faible.*

— un bassin occupé par le cône de déjection de l'Oued Nebaaha. Les courbes ont une concavité moins marquée et, certaines même, (40 et 50) s'incurvent vers le Nord. La pente hydraulique passe à 1,7 pour 1000 (au lieu de 1 au Centre).

L'étude des terrains (puits et sondages) montre que la perméabilité est plus forte. Le débit a donc augmenté et *par voie de conséquence l'alimentation.* Cet accroissement est dû aux apports de l'Oued Nebaana d'une part dont l'underflow important vient s'ajouter aux crues qui s'infiltrent rapidement sur la vaste aire d'épandage que représente son cône de déjection ; d'autre part au déversement des nappes des grès oligocènes du monoclinal de Sbikha. Ceci est mis en évidence par les courbes chimiques, courbes d'égale salure, qui s'infléchissent vers le S.E.

— un bassin d'évaporation, la plaine d'El Metbasta. La pente hydraulique diminue et devient sensiblement parallèle à la surface topographique. Ici apparaît le rôle de l'évaporation. Les courbes 40 et 35 marquent une indentation notable vers l'Ouest montrant l'absence d'apports issus des contreforts occidentaux. La pente hydraulique est faible, 0,5. Les puits montrent le colmatage progressif des sédiments. *L'écoulement est lent.* On remarquera que la courbe isopiézométrique 35 est parallèle à la courbe de niveau 40. Les terres de surface deviennent salées.

Nous avons remarqué déjà que la perte de charge augmentait d'Est en Ouest où les courbes sont parallèles aux affleurements du monoclinal de Sbikha et s'infléchissent vers le Sud. C'est très net, en particulier pour la courbe 40. Cette bordure joue ainsi un grand rôle dans l'alimentation indirecte de la nappe, rôle mis en évidence, ici encore, par l'analyse chimique des eaux.

A l'Est, au contraire, les courbes se dirigent vers le N. tangentiellement à la colline du Draa es Souatir. Il y a une importante perte de charge due au drainage de l'Oued Bassass - El Ataf qui se jette dans le lac Kelbia. En effet les fuites ne peuvent avoir lieu par le Draa es Saoutir, car ce dernier est limité à l'Ouest par une faille formant écran. D'ailleurs l'étude des courbes de la nappe du Golfe d'Enfidaville ne traduit aucune influence latérale à l'Ouest.

2) — *Secteur central.* L'unité hydraulique est constituée par les cônes de déjection des oueds Zéroud et Merguellil. L'écoulement s'opère d'une part dans le sens Sud-Ouest Nord-Est vers la cuvette de Mesbasta, d'autre part dans le sens Ouest-Est vers un accident tectonique d'orientation Nord-Sud jalonné par les sources d'Aïn el Ksar Aïn Tarfaïa. Les courbes isopiézométriques indiquent une chute rapide du niveau statique (courbe 40 et 50). *Un seuil hydraulique* retient les eaux du secteur central. Les eaux souterraines le franchissent partiellement et sont reprises par l'évaporation le long de la ligne des sources, et drainées souterrainement vers les sebkhas.

L'écoulement des nappes s'effectue suivant des courants principaux nettement marqués correspondant au lit des oueds. Les courbes marquent à leur passage une convexité tournée vers l'aval, indiquant une alimentation par infiltration des eaux des Oueds et de leur zone de débordement.

Selon le cours de l'Oued Merguellil, les courbes présentent une indentation remarquable vers l'aval avec une pente forte, voisine de 3, qui s'affaiblit ensuite, atteint l'unité pour s'accentuer avec une valeur égale à 5 vers la terminaison Nord Ouest de la dune de Draa Temmar. Cette anomalie très nette pour les courbes 35 et 40 semble due aux infiltrations des crues de l'Oued Mergueillil et à l'alimentation propre de la dune.

La perméabilité des sédiments aquifères est très forte et le débit des nappes est important.

3) — *Secteur Sud.* Les nappes de cette région sont constituées par des sédiments fins et peu perméables en continuité hydraulique avec celles du secteur précédent.

Elles sont drainées par les zones d'évaporation de l'accident Nord Sud Aïn el Ksar Aïn Tarfaïa et par les Sebkhas. Il s'agit de bassins d'évaporation clos, comme le montre la fermeture des courbes 40 et 50.

L'examen des courbes dans la partie Nord Ouest montre une contribution de l'Oued Zeroud à l'alimentation.

En résumé. L'examen des courbes izopiézométriques, en tenant compte du résultat des sondages et des analyses chimiques permet de se rendre compte du comportement des eaux souterraines dans la plaine étudiée qui contient une des réserves les plus importantes de la Tunisie.

B) — *Golfe d'Enfidaville — Lac Kelbia.*

La nappe souterraine du Golfe d'Enfidaville s'écoule du lac Kelbia vers la mer. La pente hydraulique moyenne est faible. La perte de charge diminue de la bordure S.E. (1,6) vers le Centre (1, 2) et le N.O. (0,9). Elle passe de 1,6 au N. E. du Lac Kelbia à 0,3 vers Sidi Bou Ali. Le niveau s'abaisse ainsi de la cote 25 au Sud du Lac, vers le niveau de la mer. L'observation des courbes isopiézométriques et de la pente montre donc l'existence d'un *courant privilégié* à la bordure S.E. où la nappe s'écoule vers le golfe d'Hammamet en direction de Hergla. L'alimentation latérale du Draa es Souatir ne devient notable qu'au N.O. où les courbes sont tangentes aux affleurements.

C) — *Plaine de Grombalia.*

La plaine de Grombalia, résultant du comblement d'un ancien golfe quaternaire constitue une entité hydraulique nettement individualisée. Bordée à l'Ouest par les Avants-Monts de la Dorsale Tunisienne, à l'Est par les plis du Cap Bon, au Sud par des collines miocènes, elle s'ouvre largement vers la mer au Nord. L'altitude s'abaisse progressivement du Sud-Est vers le Nord-Ouest de 150 m. au niveau de la mer. L'hydrogéologie bien étudiée par E. BERKALOFF et H. SCHOELLER et précisée par de nombreux forages permet aujourd'hui d'établir une synthèse complète. Les coupes relevées dans les puits et les sondages montrent nettement que les sédiments quaternaires aquifères perméables aux bordures Sud, Ouest et Est se colmatent vers le Centre et le Nord de la Cuvette.

L'examen des courbes isopiézométriques nous révèle une surface libre de la nappe simple et homogène dans son ensemble. Les courbes rapprochées en bordure s'espacent vers le centre et vers le Nord.

La pente hydraulique voisine de 3, *au centre s'accentue vers les rives Ouest et Est pour atteindre respectivement les valeurs* 5,8 *et* 5. Les études géologiques nous ont montré que, parallèlement, la *perméabilité allait en croissant* vers les mêmes limites. Les aires latérales de la plaine sont donc des *zones de circulation et d'alimentation.* Ce mode d'écoulement se traduit par une forme générale en *croissant* des courbes isopiézométriques.

II. — *Problèmes soulevés par les accidents tectoniques occasionnant des chutes brusques du niveau hydrostatique. Seuils hydrauliques.*

Les géologues de Tunisie ont remarqué depuis longtemps la présence d'anomalies d'un type particulier. Dans certaines structures hydrauliques on observe une chute brusque du niveau hydrostatique, chute que l'on ne peut attribuer à des variations de perméabilité ou de débit et qui sont en relation étroite avec la structure profonde. Nous les désignerons par le terme de *seuils hydrauliques.*

Il fallut attendre ces dernières années pour résoudre ce problème, solution qui, du point de vue de l'exploitation des nappes a des répercussions importantes.

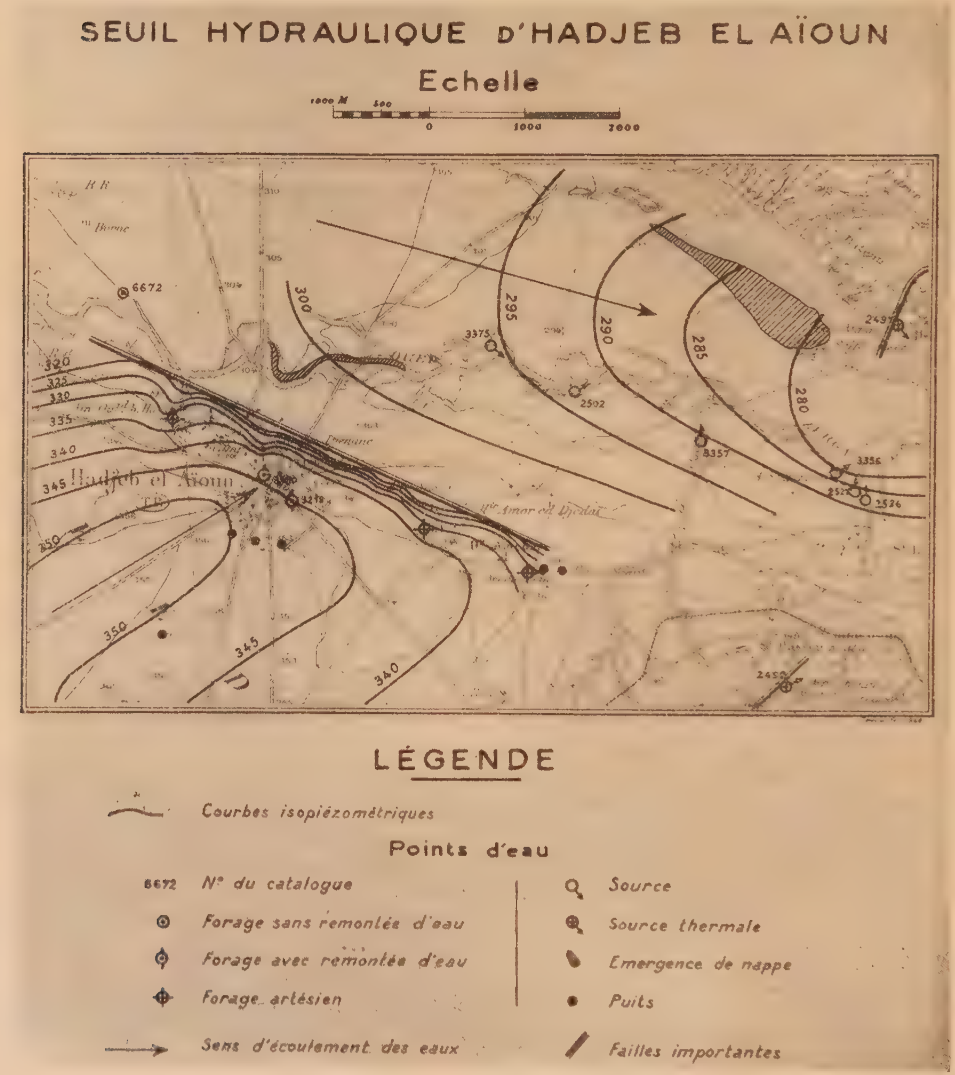


Fig. 2

L'exemple le plus typique est réalisé à Hadjeb el Aïoun. Des anomalies semblables ont été étudiées en de nombreux points de la Tunisie : Gafsa, Djilma (Bled er Roua), Aïn Tarfaïa, el Aouareb, Sbeïtla, Sbiba, Kasserine, Oued Leben (aval de Maknassy) Fériana.

Seuil hydraulique d'Hadjeb el Aïoun (*S.O. de Kairouan*) (fig. 2).

Les études de H. SCHOELLER et J. DECROCQ ont permis de tracer avec exactitude les courbes isopiézométriques de la nappe d'Hadjeb el Aïoun. Leur observation montre une structure hydraulique complexe. On peut distinguer deux compartiments séparés par la falaise quaternaire d'Hadjeb el Aïoun, orientées N.N.O — S.S.E, haute d'une trentaine de mètres et longue de douze kilomètres.

Au Sud, la région présente une structure synclinale, synclinal d'Hadjeb el Aïoun dont l'axe, de direction sensiblement S.S.O-N.N.E., passe légèrement à l'Ouest du village. Sa cote s'abaisse progressivement vers le sommet de la falaise à 340 mètres puis à 310-320 mètres au pied de cette dernière. Les courbes isopiézométriques montrent un *courant général dirigé S.S.O.-N.N.E.* avec alimentation latérale importante. Puis brusquement sur l'emplacement de la falaise, on constate une brusque chute de niveau de la surface libre de la nappe avec une perte de charge de 70 millimètres par mètre, valeur considérable que ne peuvent expliquer des variations de perméabilité ou de débit. Le pied de la pente qui correspond à la cote 310-320 est marqué par une ligne de nombreuses sources (plus de 24), à débit total moyen 30 litres par seconde, naissant par griffons et dont la plupart présentent une température plus élevée que la normale (25° centigrades). Leur teneur en résidu sec varie de 692 à 1775 milligrammes par litre.

Au Nord, s'étend la vaste vallée de l'Oued Zéroud et de son affluent l'O. Zerga. L'écoulement de la nappe change nettement de direction et *devient N.N.O.-S.S.E.* La pente hydraulique est peu accentuée, 3,6 environ. La qualité de l'eau est moins bonne, les teneurs en résidu sec sont de 2620 à 3435 milligrammes par litre.

Cette anomalie resta longtemps inexpliquée. En 1943, J. DECROCQ se basant sur la répartition des sources, leur venue par griffons et leur température, indiquant nettement une origine profonde, admit l'hypothèse d'une faille affectant le substratum et se traduisant en surface par la falaise. Les nombreux sondages de recherches hydrauliques entrepris par la Direction des Travaux Publics au cours de ces dernières années ont confirmé cette hypothèse et résolu le problème d'Hadjeb el Aïoun.

Dans le secteur Sud, 14 sondages ont recoupé des alternances de grès plus ou moins sableux et de marnes d'âge helvétien et tortonien donnant naissance à une série de *nappes captives empilées* dont le niveau statique oscille entre 345 et 350 mètres. La cote du terrain s'abaissant de 340 à 320-310 mètres, les forages situés à mi-hauteur dans les ravins transversaux sont artésiens et donnent d'excellents débits variant de 5 à 30 litres par seconde. Le résidu sec est faible de 0,5 à 2 grammes par litre. Par contre un sondage foré au Nord de la falaise a recoupé sur 418 mètres des couches à prédominance marneuse non aquifères du Miocène supérieur, montrant ainsi un décalage vertical de plus de 1.000 m., ayant affaissé la partie Nord. On est donc en présence d'une faille, ou plutôt d'une ligne de fractures résultant d'une série de décrochements d'orientation. générale N.N.O.-S.S.E. et située à la rupture de pente. L'ensemble se complique d'ailleurs de petites failles obliques ou orthogonales. Ces accidents, du point de vue lithologique, délimitent deux compartiments, l'un méridional perméable venant buter au Nord contre un complexe imperméable. Des prospections géophysiques récentes ont confirmé cette structure.

Du point de vue hydraulique l'anomalie des courbes isopiézométriques correspond donc à une faille du substratum profond assurant le colmatage du synclinal aquifère d'Habjeb el Aïoun et l'isolant presque totalement du compartiment aval hydrauliquement plus bas. Ces structures ont été désignées par J. ARCHAMBAULT sous le terme de *seuils tectoniques colmatés.* Nous leur réserverons l'appellation de *seuils hydrauliques.*

Les seuils hydrauliques apparaissent sur les cartes hydrologiques par des courbes isopiézométriques brusquement rapprochées, marquant une chute rapide de la pente de la surface libre de la nappe et par des alignements de sources artésiennes ou semi-artésiennes à température plus élevée que la normale. Cette structure implique un mode d'exploitation rationnel dont le type a été réalisé à Hadjeb el Aïoun. On a recours aux forages dans le compartiment surélevé où le niveau de la nappe est le plus haut. Compartiment déterminé par la cote des courbes isopiézo-

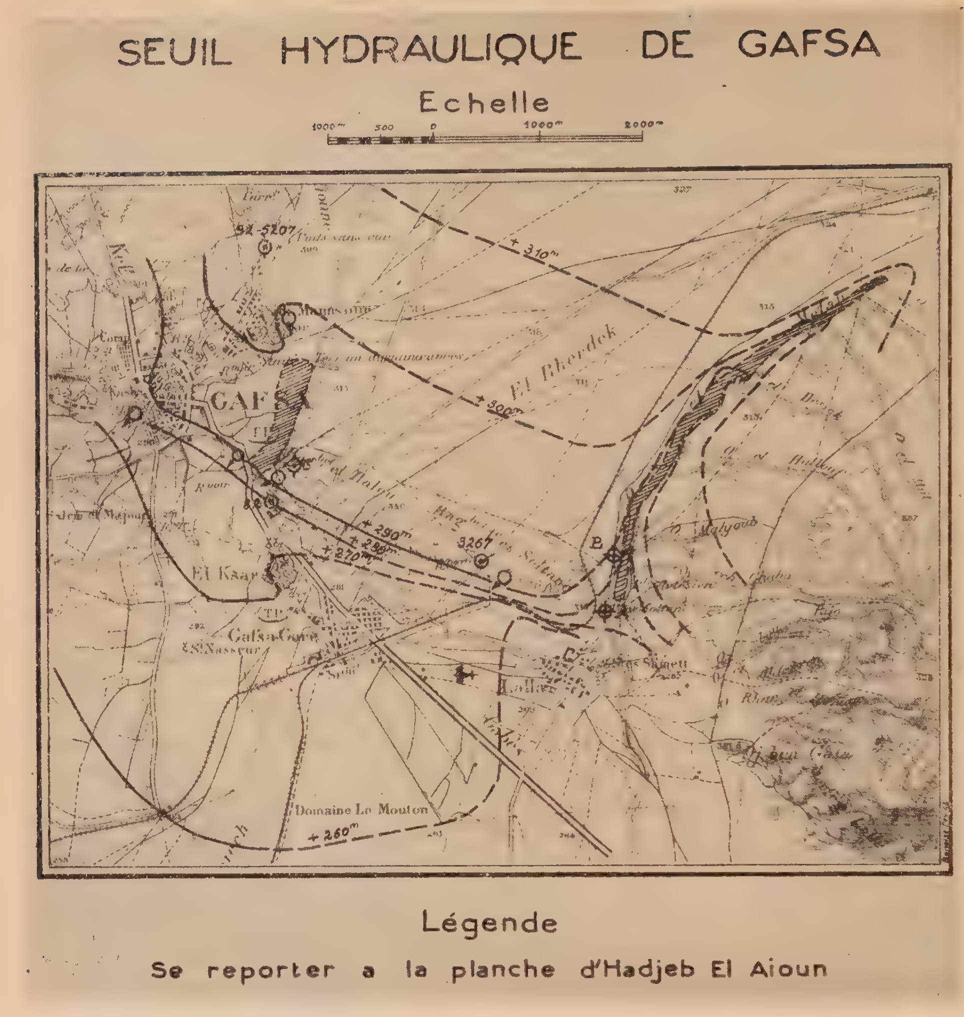


Fig. 3

métriques dont tout l'intérêt pratique apparait- ici. Pour obtenir des puits artésiens on fore dans les ravins transversaux afin de se situer à une altitude moins élevée et en amont de la faille.

Seuil hydraulique de Gafsa. (fig. 3).

La trouée de Gafsa est fermée par une falaise qui, d'El Ksar à Lalla, s'étend sur une longueur de 3 kilomètres 500 et marque une dénivellation d'une trentaine de mètres entre le plateau d'El Rherdek et la dépression de Gafsa-Gare. Accident jalonné par une série de sources à fort débit. On doit à E. BERKALOFF la carte en courbes isopiézométriques. Comme à Hadjeb on peut distinguer nettement deux compartiments hydrogéologiques; une cuvette septentrionale remplie de sables et argiles pontiennes recouvrant des calcaires très fissurés formant *drain profond* venant buter contre la plaine effondrée de Gafsa-Gare.

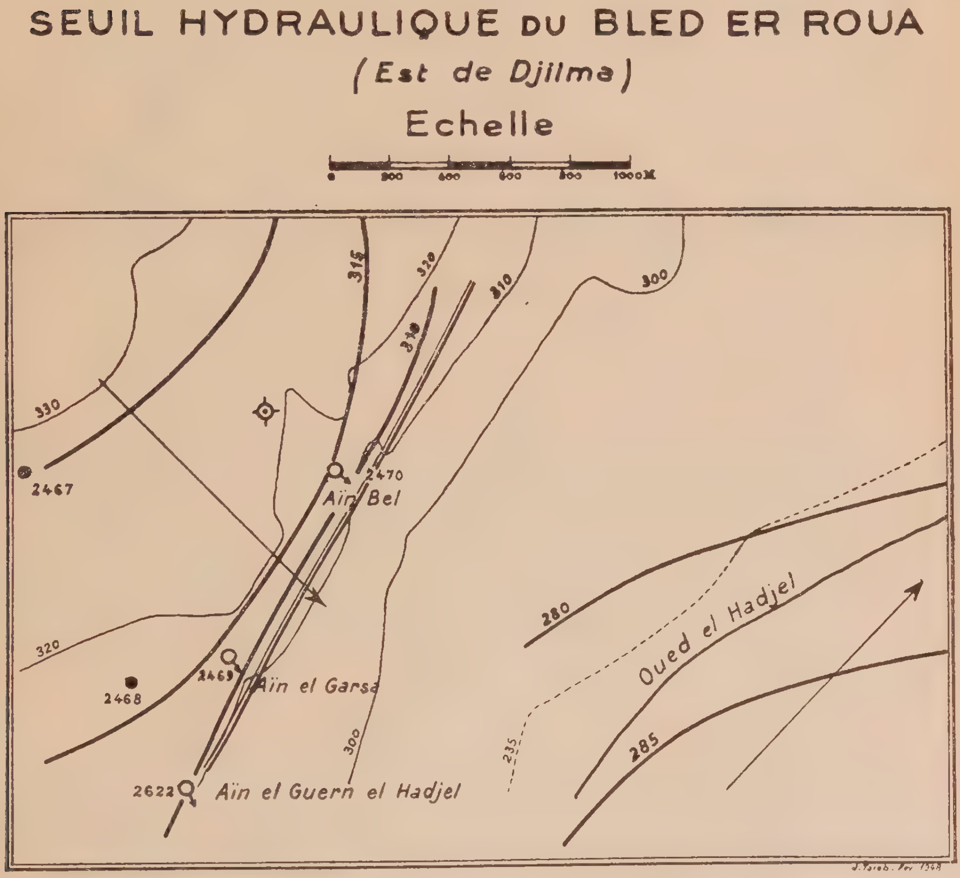
Au Nord les courbes montrent que la nappe s'abaisse lentement vers le Sud selon une pente hydraulique de 6 millièmes jusqu'à la cote 300 sur l'emplacement de la falaise. Puis brusquement le « gradient hydraulique « augmente considérablement à 70 en moyenne, la surface libre de la nappe passant à 270 mètres. Les deux points bas de cette partie Nord correspondant aux vallées des Oueds Baiech à l'Ouest et El Melah à l'Est marquent par l'inflexion prononcée des courbes vers l'amont un drainage intense de la nappe, plus important pour le second. Des sondages implantés dans ce compartiment à niveau hydrostatique élevé ont donné comme on devait s'y attendre d'importants débits. Ce sont les forages de Sidi-Mansour (5207) qui dans les calcaires cénomaniens a donné un débit spécifique très élevé et les puits artésiens de Lalla (A. et B.).

Au Sud, le niveau de la surface libre de la nappe s'abaisse lentement vers le Sud-Est ($i = 3$ millièmes) où il s'équilibre avec les émergences du Djérid (Tozeur-Nefta). Les sondages et les recherches géophysiques y ont montré l'enfouissement des nappes profondes et l'effondrement des calcaires.

La falaise de Gafsa correspond donc à une faille donnant naissance à un *seuil hydraulique.*

Cette conclusion a été vérifiée récemment par le sondage d'El Ksar (C). Implanté au Nord du tracé supposé de la faille, il a donné un débit artésien de 105 litres par seconde au niveau du sol et a vérifié l'affaissement des calcaires vers le Sud.

Seuil hydraulique de Djilma (*Bled Er Roua*). (fig. 4)



Pour la légende se référer à la planche d'Hadjeb el Aïoun

FIG. 4

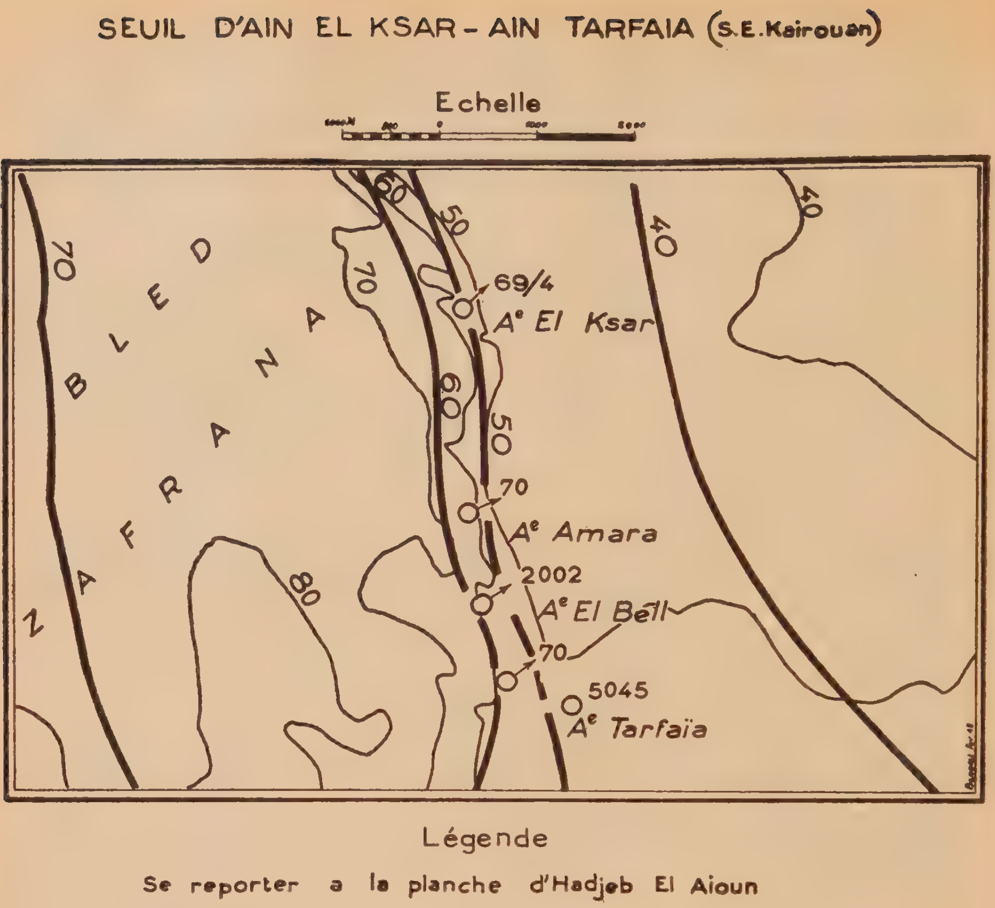


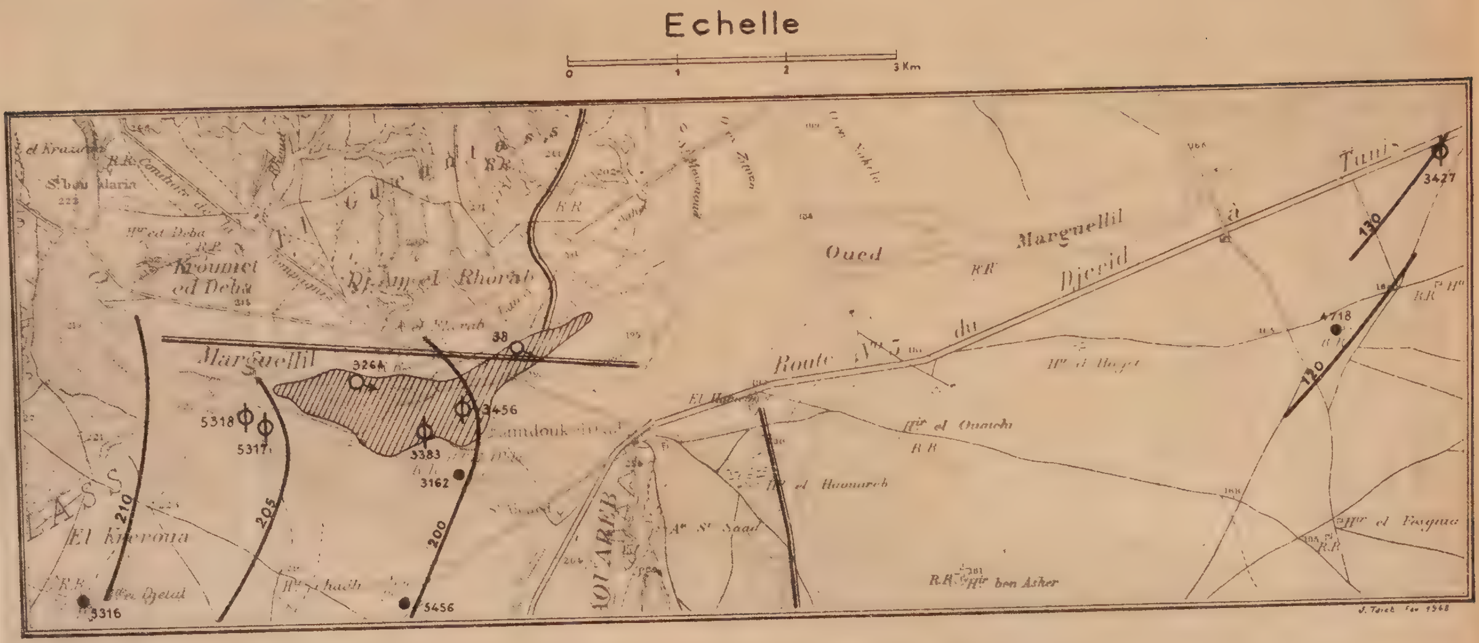
Fig. 5

Au Bled Er Roua à l'Est de Djilma (Tunisie Centrale) la carte des courbes isopiézométriques marque une importante anomalie du type *seuil hydraulique.* Sur l'emplacement d'une falaise topographique de direction Nord-Sud correspondant à une dénivellation d'une quinzaine de mètres, les deux courbes isopiézométriques 310 et 315 se rapprochent brusquement, marquant une chute rapide du niveau de la surface libre de la nappe. Selon cet alignement naissent trois sources dont la plus méridionale l'Aïn Guern El Hadjel présentent d'importants griffons. Leur résidu sec est nettement moins élevé que celui des eaux des puits voisins. Cette structure nous a conduit à l'hypothèse d'une faille se traduisant en surface par la falaise. Nos conclusions ont été confirmé par le forage d'un puits sur le bord du plateau ayant rencontré des nappes captives profondes de niveau statique plus élevé que la surface du sol, et fournissant un bon débit.

Seuil hydraulique d'Aïn Tarfaia (*S. E. de Kairouan*) (fig. 5)

La carte des courbes isopiézométriques de la cuvette des Sebkras Sidi el Hani-Chérita marque, au S. O. de Kairouan une importante anomalie. Au pied d'une falaise topographique d'une dizaine de mètres orientée Nord-Sud, sur 5 kilomètres de long, s'alignent une douzaine de sources dont le débit total atteint une vingtaine de litres par seconde. Au passage de cet accident les deux courbes isopiézométriques 40 et 50 se rapprochent nettement indiquant ainsi une perte de charge importante. Il est vraisemblable que nous nous trouvons ici en présence d'un accident tectonique analogue à ceux décrits précédemment. Des recherches hydrauliques et géophysiques en cours viendront résoudre ce problème.

SEUIL HYDRAULIQUE D'EL AOUAREB



Légende – Se reporter à la planche d'Hadjeb el Aïoun

FIG. 6

Seuil hydraulique d'El Aouareb (*S. O. de* Kairouan) (fig. 6)

Avant d'atteindre la plaine quaternaire de Kairouan, l'Oued Marguellil franchit un col séparant la terminaison septentrionale du Djebel el Aouareb du Djebel Chérichira. Le Djebel el Aouareb apparaît comme un pli-faille à regard oriental dont le flanc inverse est effondré sous la plaine de Kairouan. Il en est de même pour le Djebel Chérichira. Du point de vue hydraulique on peut considérer deux compartiments, l'un à cote élevée à l'Ouest et, à l'Est, la plaine effondrée de Kairouan.

A l'Ouest, le cours de l'Oued Marguellil présente un débit pérenne important. Des sondages hydrauliques permettent d'en connaître la structure profonde. Le plus important (5318) placé rive droite sur les alluvions dans le prolongement des couches du Miocène supérieur du Djebel Ain el Rhorab (terminaison méridionale du Djebel Chérichira) à un kilomètre au plus vers le Sud a traversé 177 mètres d'alternances de sables, de cailloutis et de marnes attribuables au Quaternaire puis les couches de l'Eocène moyen sur 273 mètres. Il y a donc une faille secondaire Est-Ouest ayant causé l'effondrement du compartiment de la vallée de l'Oued Marguellil. Il a recoupé une série de nappes empilées dont le niveau de la surface libre est à 3 mètres au-dessous du sol. L'eau est de bonne qualité. Des forages voisins ont confirmé le niveau de la nappe (— 3,50).

A l'Ouest dès le franchissement de l'extrémité septentrionale du Djebel el Aouareb, les eaux de l'Oued s'infiltrent dans le sol, et les points d'eau de la plaine de Kairouan montrent que les niveaux de ce compartiment sont très profonds, (40 mètres en moyenne) et nettement inférieurs.

La région d'El Aouareb correspond donc à un *seuil hydraulique*, résultant des fractures importantes orientées sensiblement N.S. et qui se traduisent latéralement par les pli-failles des Djebels Chérichira et El Aouareb.

Les quelques exemples ci-dessus montrent l'importance des *seuils hydrauliques* dans l'hydrogéologie de la Tunisie. Ce sont des points de captage privilégiés. Pour les interpréter correctement en vue de l'exploitation l'étude de la topographie et de la géologie de surface est souvent insuffisante et doit être complétée par des observations hydrauliques, des sondages et des mesures de résistivité électrique qui décèlent des failles difficilement apparentes.

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4) VARIATIONS DE LA HAUTEUR PIEZOMETRIQUE DANS LES NAPPES ARTESIENNES

par

L. J. TISON, Professeur à l'Université de Gand, Directeur du laboratoire d'hydraulique.

1) Des sondages assez nombreux ont été réalisés le long du tracé d'une coupure qui doit relier l'Escaut supérieur à l'Escaut Maritime, en contournant la ville de Gand.

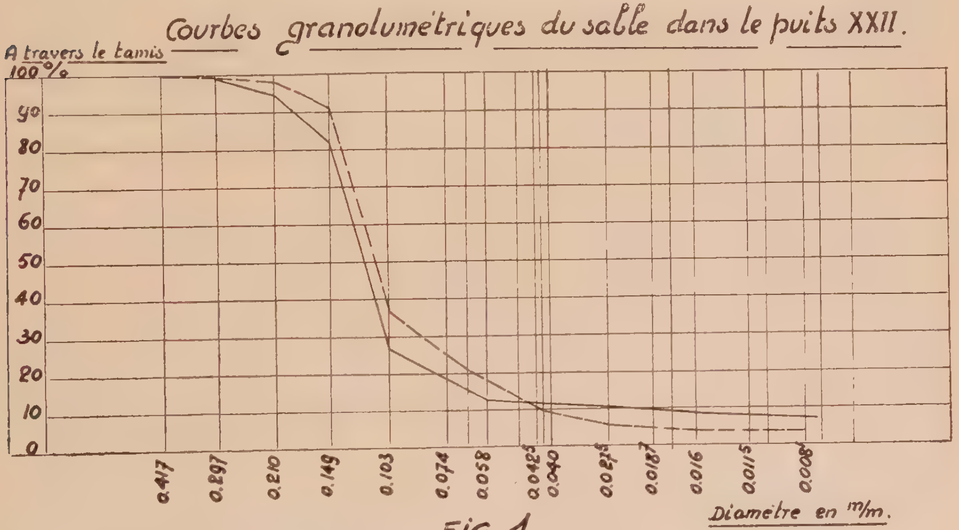
Ces sondages ont rencontré les couches suivantes, assez sensiblement horizontales sur une certaine étendue :

a) une couche supérieure de sable fin d'environ 7.50 m. d'épaisseur;

b) une couche d'argile tertiaire de 4 à 5 mètres d'épaisseur;

c) sous l'argile, une couche sable de 10.15 m. d'épaisseur moyenne. Cette couche de sable est limitée vers le bas par une couche gréseuse peu épaisse et cependant de perméabilité assez réduite. La nappe dans la dernière couche de sable mentionnée est artésienne, mais elle est en communication avec l'Escaut vers le Nord, tout en s'étendant assez loin vers le sud-est en ne devenant localement libre que dans la traversée de quelques vallées.

2) Un pompage d'essai fut réalisé en avril-mai-juin 1946, dans cette nappe artésienne en vue d'en déterminer les caractéristiques. Un puits, pour le pompage, de 0.20 m. de diamètre, fut descendu jusqu'à la couche gréseuse. Par suite de la finesse du sable, le puits fut entouré d'une couche de fin gravier, puis d'une couche de sable grossier. La figure 1 donne la composition granulométrique de deux échantillons du sable porteur de la nappe artésienne.



Courbes granolumétriques du sable dans le puits XXII.

FIG. 1.

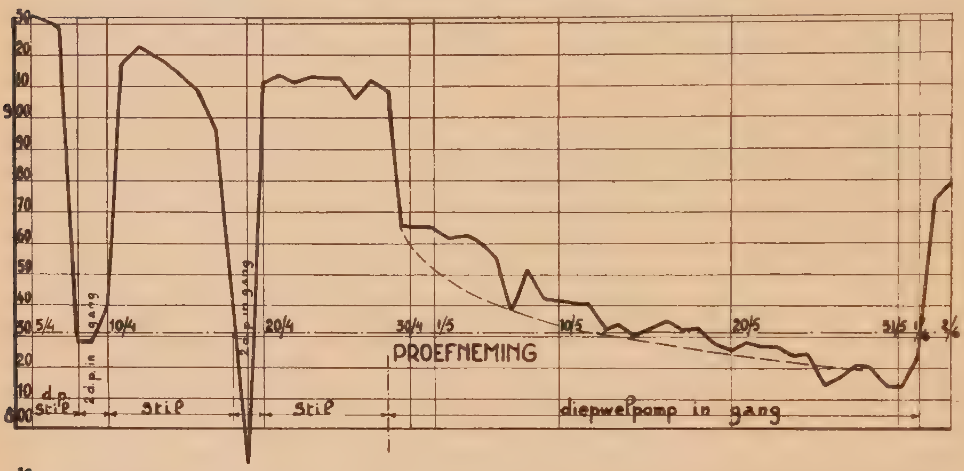
Douze puits d'observation numérotés de XIII à XXIV furent établis suivant deux directions perpendiculaires, de part et d'autre du puits de pompage, quatre d'entre eux à une distance de 20 m. de ce dernier puits (puits XIII, XVI, XIX et XXII), quatre autres à 60 m. du puits de pompage (puits XIV, XVII, XX et XXIII) et les quatre derniers à 150 m. de ce puits (puits XV, XVIII, XXI et XXIV). Ces puits étaient donc disposés par groupes de trois sur les quatre branches d'une croix.

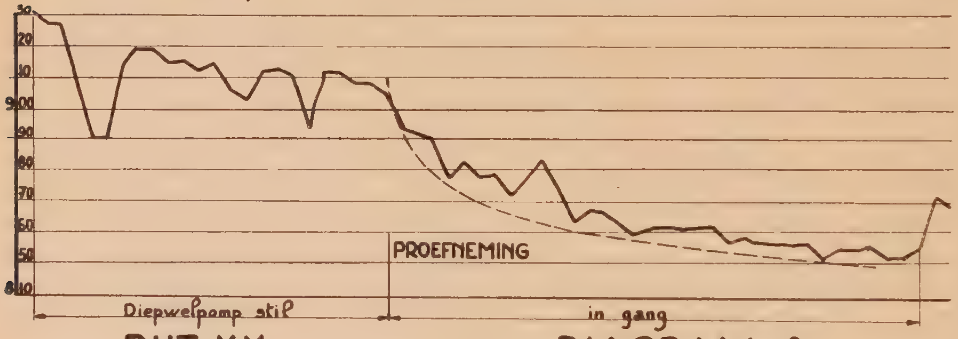
Des essais préliminaires avec un débit assez peu constant de 1 litre par seconde eurent lieu du 4 au 7 avril et du 18 au 19 avril. Mais le véritable pompage d'essai ne commença que le 29 avril et dura jusqu'au 2 juin : le débit était alors de 0.5 litre par seconde.

Les diagrammes 2, 3, 4 et 5 reproduisent les variations du niveau dans quatre de ces puits d'observation durant les essais. Bien que les essais préliminaires ne soient spécialement indiqués que sur le premier de ces diagrammes (relatif au puits XIX, situé à 20 m. du puits de pompage), il a lieu d'observer que ces premiers pompages exercèrent bien leur influence sur les autres puits d'observation, comme on peut d'ailleurs s'en rendre compte. En d'autres termes, c'est le diagramme 2 relatif au puits XIX qui donne seul les indications complètes et exactes au sujet des dates des pompages.

A remarquer encore que ces diagrammes ont été construits à l'aide des résultats d'observations faites de 24 en 24 heures. C'est ainsi que pour les pompages préliminaires certains maxima de dépression n'ont pu être observés ; il n'en est pas de même pour le pompage définitif et continu du mois de mai.

PUT XIX DIAGRAM 2

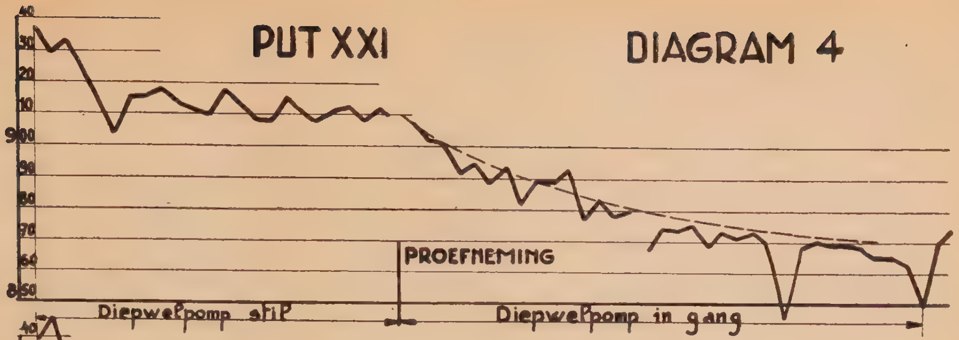




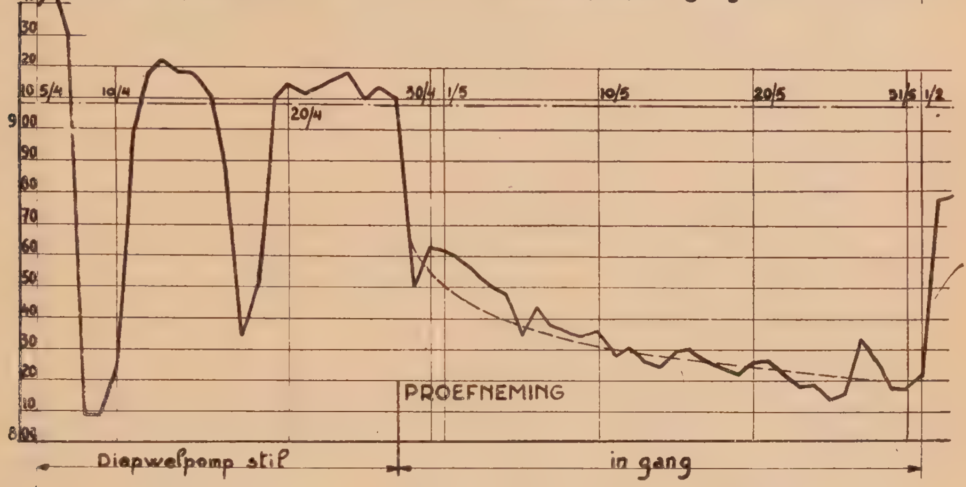
PUT XX DIAGRAM 3

D'autre part, il sera question plus loin des résultats d'observations continues sur divers puits pendant des périodes assez courtes.

3) Le coefficient de perméabilité $\frac{m'}{\mu}$, fut déterminé en supposant que la situation au 31 mai 1946 était une situation permanente, d'équilibre, ce qui était d'ailleurs sensiblement le cas pour les douze puits d'observation dans lesquels le niveau ne variait plus guère en fonction du temps. En prenant les niveaux dans deux puits quel-



PUT XXI DIAGRAM 4



PUT XXII DIAGRAM 5

conques d'une des branches de la croix formée par la situation des puits en plan, nous avons déterminé 12 valeurs de $\frac{m'}{\mu'}$ à l'aide de la relation :

$$Q \, ln \frac{R_2}{R_1} = \frac{m'}{\mu'} 2 \pi e (h_2 - h_1)$$

Q est le débit 0.0005 m^3/s, R_1 tet R_2 sont les distances au centre du puits de pompage de deux puits d'observation quelconques dont les hauteurs d'eau par rapport à la couche gréseuse sont h_1 et h_2; e est l'épaisseur de la couche artésienne. On a trouvé $\frac{m'}{\mu'} = 3.10^{-5} \, m/s$.

4) Nous nous sommes alors proposé d'étudier la descente de la hauteur piézométrique dans les différents puits d'observation en fonction du temps à l'aide des diverses théories présentées et de comparer les résultats de ces études avec ce qui a été observé.

a) Nous avons d'abord utilisé une méthode proposée par Weber (1) et qui est basée sur une hypothèse suivant laquelle le débit Q_r qui traverse un cylindre de rayon r est donné par :

$$Q_r = Q \left(1 - \frac{r^n}{R^n}\right)$$

n étant sensiblement égal à 1,6 d'après Weber; R est la distance du centre du puits de pompage au bord de la dépression provoquée par ce pompage.

La loi de Darcy appliquée à un puits artésien et tenant compte de l'hypothèse précédente conduit à :

$$Q = \frac{2\pi e \frac{m'}{\mu'} (H - h)}{ln \frac{R}{r} - \frac{1}{n} \frac{R^n - r^n}{R^n}} \qquad (1)$$

H étant la hauteur de l'eau dans un puits d'observation avant le pompage et h étant cette hauteur après pompage, à une distance r.

Dans le cas d'une nappe artésienne, la réduction de pression due au pompage n'est pas nécessairement accompagnée de la formation d'un entonnoir dans la nappe. Une réduction ΔV d'un volume V étant occassionnée par une réduction de pression $-\Delta p$ due à une différence $H - h$ de la hauteur de la colonne liquide, on a :

$\Delta V = - V \frac{\Delta p}{C}$, C valant 50.000.000 environ si p est exprimé en atmosphères.

On en déduit aisément que $dV = \frac{\pi m'}{5 . C} . e . (H - h) r . dr$ (2)

qui conduit, compte tenu de (1) à :

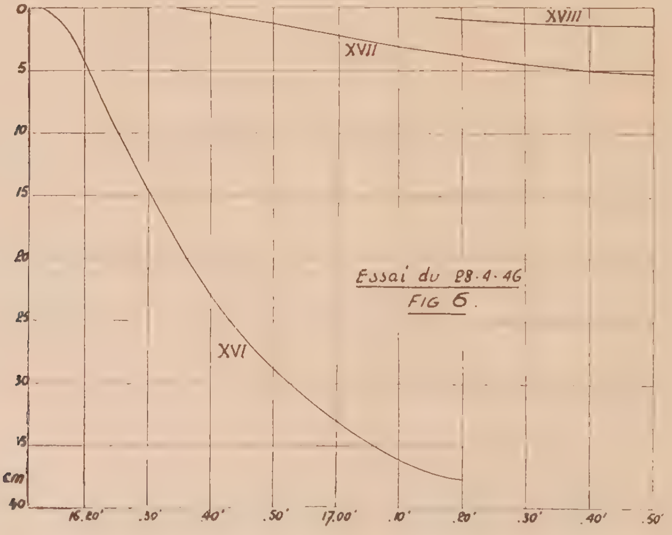
$V = \frac{Q\mu'}{10 . C} R^2 \frac{n}{4(n + 2)}$, moyennant certaines approximations.

Ce volume V est le volume d'eau qui est libéré et parvient au puits de pompage quand la réduction de pression s'est transmise dans un cylindre de rayon R.

On peut en déduire que $R = 67.000 \sqrt{\frac{T}{\mu'}}$ qui avec l'aide de l'équation (1) permet de calculer h pour un r et un T donnés.

Nous avons fait d'assez nombreuses applications de cette théorie au problème exposé ci-dessus. Les résultats sont assez concordants avec ceux de l'observation pour autant que l'on prenne pour valeur du rayon R du champ influencé par le pompage, celle qui a été observée.

Ainsi pour le début de l'essai du 26 avril 1946, la variation du niveau en fonction du temps, relevé de façon continue à l'aide de limnigraphes, dans les puits XVI, XVII et XVIII est donnée par le diagramme 6. Quand le niveau commence à baisser au puits XVII, R vaut évidemment 60 m., écartement de ce puits du puits de pompage. Pour le puits XVI, r vaut 20 m. et il y correspond d'après la théorie de Weber



Essai du 28.4.46

FIG 6.

une baisse du niveau dans ce puits égale à :

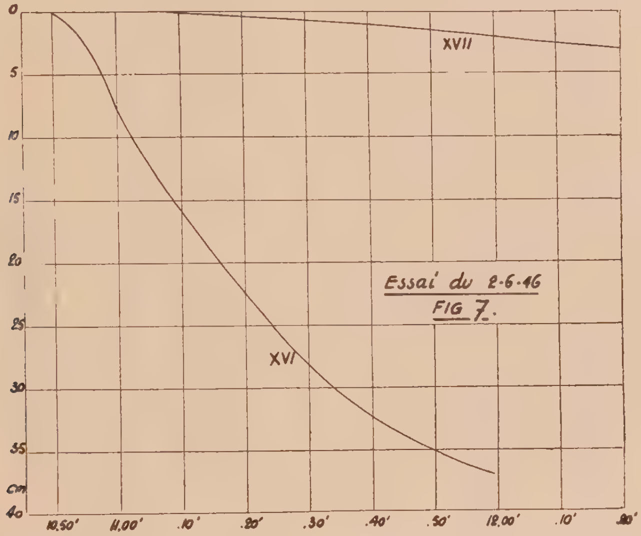
$$H - h = \frac{Q\left(ln\frac{R}{r} - \frac{1}{n}\frac{R^n - r^n}{R^n}\right)}{2.\pi.e\frac{m'}{\mu'}}$$

soit 0.14 m. au lieu de 0.19 observé.

Quand le niveau commence à baisser dans le puits XVIII, R vaut 150 m., on doit trouver dans le puits XVI une baisse H — *h* donnée par la formule ci-dessus, soit 0.363 m. au lieu de 0.364 m. observé.

Au même moment dans le puits XVII (*r* vaut 60 m.), H — *h* calculé vaut 0.10 m. au lieu de 0.04 m. constaté.

Lors d'un essai du 2 juin 1946, les variations de niveau constatées dans les puits XVI et XVII sont reproduites par le diagramme 7. Quand le niveau commence à baisser dans le puits XVII (R vaut 60 m.), le niveau calculé de XVI vaut 15.5 cm. au lieu de 14 observé.



Essai du 2-6-46

FIG 7.

La concordance est par conséquent assez bonne en général. Mais par contre, quand on fait intervenir dans les calculs la valeur de R (rayon du cone d'action du puits de pompage à un instant donné) fournie par la formule 87.000 $\sqrt{\frac{T}{'\mu}}$, il ne peut plus être question d'une concordance même assez grossière. Ainsi, pour un T valant 24 minutes, R serait égal d'après cette formule à 28150 m. et la descente du niveau au puits d'observation situé à 20 m. serait de 1.90 m. alors qu'il ne fut constaté que 0.19 m. !

De même, pour T valant 40 jours, R vaudrait 1.390.000 m. d'après la formule en question et conduirait à un H — *h* égal à 3 mètres alors qu'on n'a même pas constaté 1 mètre.

Il en résulte que cette théorie de Weber donne pour les nappes artésiennes des résultats qui ne peuvent être pris en considération dès qu'ils sont basés sur le calcul de R.

b) Nous avons eu connaissance après guerre des travaux de Muskat, Meinzer, Theiss et Jacob (2) et nous avons voulu appliquer leurs résultats et particulièrement ceux de Jacob au problème qui nous intéressait. La théorie de Jacob n'introduit aucune hypothèse autre que celle de Darcy et permet de résoudre complètement le problème. Elle est basée sur le fait que la variation de pression due au pompage produit non seulement une expansion du liquide comme il en est tenu compte ci-dessus, mais aussi une variation des vides du « squelette « constitué par les grains de la couche perméable. Les calculs de Jacob le conduisent à la considération de divers facteurs :

$$c \text{ donné par } \frac{Q}{4\,\pi \dfrac{m'}{\mu'}\, e}, \qquad b \text{ donné par } \frac{r^2\,S}{4 \dfrac{m'}{\mu'}\, e}$$

S étant un coefficient d'emmagasinement.

Jacob trouve alors une relation entre h/c et t/b, h étant la baisse de la hauteur piézométrique d'un puits pour un temps t compté à partir de l'origine du pompage. Cette relation est la suivante :

$$h\text{–}c = -0.5772 + ln\, t/b - b/t + (b/t)^2/2.2\,! - (b/t)^3/3.3\,! + (b/t)^4/4.4\,! - \ldots$$

Pour les puits d'observation situés à 20 m. du puits de pompage, on a observé que la baisse moyenne du niveau après un jour de pompage était de 0.45 m. Il en résulte que :

c vaut 0.13, de sorte que h/c est égal à 0.45/0.13 *soit* 3.45.

Le t/b correspondant d'après la relation ci-dessus est 58.

D'où : $b = \dfrac{86400}{58}$ soit 1500.

Quant à la valeur de S, elle s'écrit :

$$S = \frac{4b \dfrac{m'}{\mu'}\, e}{r^2} \quad \text{soit :} \quad \frac{4 \times 1500 \times (3 \times 10^{-5}) \times 10.15}{20^2} = 4.5 \times 10^{-3}$$

Il est alors possible de trouver pour les puits d'observation situés à 20 m. du puits de pompage, le rapport t/b pour un temps quelconque t ; la relation ci-dessus donne la valeur correspondante de h/c, d'où h.

Ces déterminations ont été faites pour 1, 5, 10, 20 et 30 jours et les résultats en ont été reportés sur les diagrammes 2 et 5, donnant la ligne pointillée qui est donc la courbe de variation calculée du niveau de l'eau dans ces puits en fonction du temps pendant le pompage d'essai. On voit que la concordance entre cette courbe calculée et la courbe observée est excellente, tant pour le puits XIX que pour le puits XXII.

Pour les puits d'observation situés à 60 m. du puits de pompage, il faut déduire la valeur correspondante de b en partant du coefficient d'emmagasinement S :

$$b = \frac{Sr^2}{4 \dfrac{m'}{\mu'}\, e} = \frac{4.5 \times 10^{-3} \times 60^2}{4 \times 3 \times 10^{-5} \times 10.15} = 13500$$

On peut alors déterminer comme ci-dessus les valeurs de h/c correspondant à des valeurs données de t et par conséquent de t/b et on en déduit les valeurs de h.

Ces déterminations ont été faites pour les mêmes temps que ci-dessus et les résultats obtenus ont été reportés sur le diagramme 3 où ils sont représentés par la ligne pointillée dont la concordance avec la courbe observée est encore très bonne.

Pour les puits d'observation situés à 150 m. du puits de pompage, la valeur de S permet de trouver que b vaut 83.000 et des calculs analogues aux précédents conduisent à la courbe calculée représentée en pointillés sur diagramme 4 pour la comparer à la courbe observée de la descente du niveau dans un de ces puits.

5) Nous désirons cependant faire quelques observations au sujet de la concordance des valeurs calculées de la baisse du niveau dans les divers puits avec les valeurs observées corresponsantes.

La comparaison en question donne une concordance excellente sur les diagrammes 2, 3, 4 et 5 et elle est tout aussi bonne pour les diagrammes analogues établis pour les autres puits d'observation.

Nous avons voulu nous rendre compte s'il en était de même pour le début du pompage. Des limigraphes inscrivant la descente du niveau dans trois puits d'observation XVI, XVII et XVIII avaient été installés avant le début du pompage définitif le 18 avril 1946. La figure 6 reproduit les courbes de descente de niveau en fonction du temps. L'origine du temps correspond avec le commencement du pompage.

Le diagramme 7 donne les courbes de descente du niveau dans les puits XVI et XVII au cours d'un nouvel essai effectué le 2 juin 1946.

Or, si nous calculons la descente du niveau dans le puits XVI (à 20 m. du puits de pompage) une heure après le commencement du pompage, on trouve : $t/b=2.4$ ce qui correspond à $h/c=0.7$ soit $h = 0.09$ m., alors que la descente observée, d'après les deux diagrammes 6 et 7, fut de 0.34 et 0.36 m. La discordance est assez nette. Elle est même relativement plus grande encore quand t est inférieur à 1 heure. Par contre, elle diminue très rapidement quand le temps augmente. Ainsi, après 6 heures de pompage, le 28 avril la descente de niveau constatée dans le puits XVI fut de 0.41 m., alors que la valeur calculée correspond à $t/b=21600/1500$ soit 14.5 qui donne $h/c = 2.3$ soit $h = 0.30$ m. Nous savons d'autre part que la concordance est quasi parfaite 1 jour après le début du pompage.

Des constatations analogues peuvent être faites pour le puits XVII situé à 60 m. du puits de pompage. Ainsi, après 6 heures de pompage, on y constate une baisse de 12.7 cm. Le calcul donne $t/b = 21.600/13.500$, soit 1.6, valeur à laquelle correspond $h/c = 0.5$ soit h égal à 0.065 m. De même, nous savons qu'après un jour de pompage, la concordance entre les valeurs calculée et observée est devenue très bonne.

Ces discordances relatives aux premières heures du pompage peut être expliquée par les considérations suivantes. Nous avons admis pour le calcul de la descente des niveaux dans les divers puits d'observation que le débit pompé était 0.0005 m^3/s Il en est bien ainsi, sauf pendant les tout premiers temps du pompage. Au début, il est assez difficile de réaliser un débit aussi faible que celui qui vient d'être indiqué. La hauteur de l'eau dans le puits de pompage ne descend qu'assez lentement et pendant toute cette période le débit de la pompe est supérieur, et parfois de beaucoup, au débit de régime. Le débit qui est pompé pendant les premiers temps est donc en moyenne très supérieur à ce débit de régime, pour lesquel les calculs sont effectués. Il n'y a par conséquent rien d'étonnant à ce que la descente du niveau dans les puits d'observation pendant les premiers temps dû pompage soit plus rapide que celle déduite des calculs. Nous avons fait des mesures du débit pompé effectivement pendant cette période de mise en train et nous avons toujours trouvé des débits supérieurs à 0.0005 m^3/s.

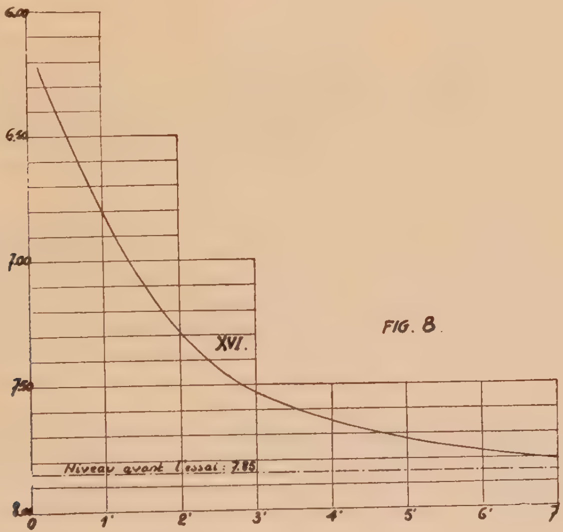
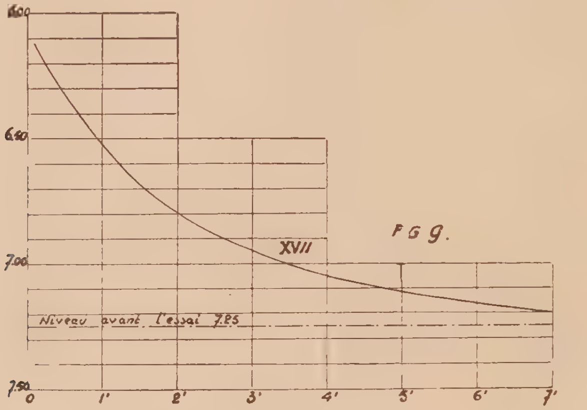


FIG. 8.

Par contre, la comparaison des courbes observées et des courbes calculées de la descente du niveau en fonction du temps dans les divers puits d'observation, montre qu'en dépit d'une bonne concordance générale, les descentes calculées sont légèrement supérieures aux descentes observées après la période initiale dont il vient d'être question ci-dessus (voir en effet les diagrammes 2, 3 et 5 surtout pour la période entre le second et le dixième jour). Nous croyons que l'explication de cette constatation est la suivante : les calculs donnent la cote de la hauteur piézométrique dans la nappe artésienne aux divers endroits où des puits d'observation ont été réalisés, mais ces calculs supposent en fait que ces puits d'observation n'existent pas.

L'existence de ces puits a pour conséquence, lors de la baisse de la hauteur piézométrique de la nappe, de provoquer une alimentation de celle-ci par l'eau du puits, ce dernier fonctionnant en fait comme un puits absorbant. La hauteur piézométrique de la nappe en est localement relevée et la hauteur de l'eau dans le puits est du fait supérieure à la hauteur piézométrique que prendrait la nappe en cet endroit si le puits n'existait pas.

Les diagrammes 8 et 9 montrent la variation de la hauteur de l'eau dans les puits en fonction du temps après qu'on y eut relevé artificiellement le niveau de 1.50 m. environ en y déversant de l'eau. Ces diagrammes ne portent malheureusement que sur les premières minutes, mais après un jour, le niveau se trouvait encore deux centimètres au-dessus du niveau initial.



D'autre part, les couches argileuse et surtout gréseuse limitant la couche étudiée ne sont pas parfaitement imperméables et une partie du débit relevé par la pompe (très faible sans doute) provenait de ces couches, réduisant ainsi d'autant le débit de la nappe artésienne proprement dite. Il doit en résulter de légers écarts entre les hauteurs observées et les hauteurs calculées, bien que la détermination du coefficient *c* à l'aide de la valeur pompée de Q ait plus ou moins affecté les résultats des calculs de cette influence.

6) Variations du niveau de l'eau dans les puits d'observation sans pompage. Des observations sur les puits avant le pompage ont montré que tandis que la pression atmosphérique passait de 750 à 770 mm. de mercure d'une façon assez continue du 1 au 17 mars, le niveau de l'eau dans les puits (dans chacun des douze puits) montait de 9 à 10cm. Du 27 mars au 4 mars, la pression barométrique restait pour ainsi dire constante et le niveau de l'eau dans les puits ne se modifiait guère.

(1) Weber. Die Reichweite von Grundwasserabsenkungen mittels Rohrbrunnen Berlin 1928.
(2) Meinzer. Compressibility and elasticity of artesian aquifers. 1928.
Muskat. M. The flow of homogeneous fluids through porous media. Mc Graw Hill. Bk. N.Y. 1937
Theis C. Relation between the lowering of the piezometric surface and the rate and duration of discharge of well using groundwater storage. *Trans. Amer. Geophys. Union.* 1935.
Jacob. On the flow of water in an elastic artesian aquifer. *Trans. Amer. Geoph. Union.* 1940.

5)

FLUCTUATIONS OF WATER LEVELS AND ARTESIAN PRESSURES IN WELLS IN THE UNITED STATES: THEIR MEASUREMENT AND INTERPRETATION [1]

by

W. F. GUYTON [2]

January 1948

Introduction

O.E. Meinzer [3] has said (1) [4] « Ground water is one of the Nation's most valuable mineral resources. It is nearly uniform in quality and coolness the year round. It bridges the gap between rains and maintains the flow of streams in dry weather.

« America spends more than a million dollars every day for ground water. The amount used daily for various purposes has been estimated as follows :

	U. S. Gallons	Cubic meters
Irrigation	10,000,000,000	38,000,000
Industrial	5,000,000,000	19,000,000
Municipal	3,000,000,000	11,000,000
Rural	2,000,000,000	8,000,000

The total use of about 20,000,000,000 gallons, or 76,000,000 cubic meters, a day is the equivalent of about 150 gallons, or 570 liters, a day for each person in the United States.

According to A.N. Sayre [5] (2) « In 1935, the total use of ground water in the United States amounted to about 10.000,000,000 gallons a day. The development was greatly accelerated by the needs of the war, so that by 1945 the total pumpage had nearly doubled... Although the war ended more than two years ago, there has been no sign of a decrease in the use of ground water. In fact, both surface water and ground water are now being used in greater quantities than ever before. As the use of water approaches ever more closely the limits of the available supply, it is believed that water-resources investigations will be needed with ever increasing urgency because water supply constitutes an important limiting factor in the continued development of the Nation's industrial and agricultural economy ».

A long-range program of technical studies, the basis for sound development and conservation of the Nation's ground-water resources, is conducted by the United States Geological Survey in cooperation with State geological surveys, State engineers, minicipalities, and water districts. About 7,000 observations wells, some 350 of them equipped with automatic water-stage recorders furnish records of the rise and decline of ground-water levels.

The measurement and interpretation of water levels in wells is an integral part of almost every ground-water investigation. Hardly a report on ground water is written that does not deal with water levels in one day or another. Recognizing this, A.M. Piper [6] wrote (3) in 1936 « Although ground-water investigations differ widely in purpose and in the degree of accuracy that is required in the base data all have one element in common : the behavior of water in natural pervious materials is deduced largely from water-level data gathered in observation wells. If the deductions are to be thustworthy, the observation well must indicate the true ground-water level within the suitable small limits ». Later Piper wrote [4] « Persons charged with the development and maintenance of adequate water supplies will be increasingly concerned with the safe yield of ground-water basins — that is, with the volume of water that can be withdrawn year after year without ultimately depleting the supply. To determine the safe yield of a basin involves an intensive study of the fluctuations in ground-water level... discriminating and evaluating each type of the fluctuation with respect to replenishment of and withdrawal from the basin. »

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[1] Approved for publication by the Director of the U. S. Geological Survey.

[2] Hydraulic Engineer, U. S. Geological Survey.

[3] President of Commission on Subterranean Water, International Association of Scientific Hydrology. Retired Geologist in charge, Division of Ground Water, Water Resources Branch, U. S. Geological Survey.

[4] See bibliography at end of paper.

[5] Geologist in charge, Division of Ground Water, Water Resources Branch, U. S. Geological Survey.

[6] Staff Scientist, U. S. Geological Survey.

In 1935 Meinzer reported [5] « Most of the observation wells have been in areas of heavy withdrawals of water from wells by pumping or artesian flow, and the measurements have been carried on in connection with the more intensive quantitative investigations to determine the safe yield of ground water in these areas. The final answer as to the quantity of water available for recovery is seldom accurately and positively obtained at the completion of an intensive investigaton covering a few years. It is therefore highly desirable that after the completion of such an intensive investigation a program be continued for obtaining records of the water levels in a number of key wells... and of obtaining records of pumpage or artesian flow in the areas in which the wells are situated. Moreover, there are many areas in the United States where no intensive ground-water investigations have been made. In the future there will be need for studies in these areas, and records of past water-level fluctuations and withdrawals will be exceedingly valuable when such investigations are undertaken. »

Recently Sayre wrote (6) « The present water-level program of the Division of Ground Water [U.S. Geological Survey] was initiated in 1934 [and later expanded] largely because of the public interest in water levels resulting from the droughts of 1930, 1934, and 1936. The program... has been carried on in connection with cooperative investigations. The wells were selected from wells that were being, observed in local or regional investigations or from state-wide observation-well programs. The water-level measurements have been made and tabulated and brief discussions have been written by the men in the field offices, and the reports are assembled and prepared in Washington for publication as annual Water-Supply Papers ».

In this outline (7) for preparing contributions to these annual water-level reports L.K. Wenzel [7] requested that the following information be included, in addition to the water-level measurements : *Descriptive text* —. Statement regarding purpose and status of cooperative investigation through which the records of water level are collected; general geography, geology, and water resources of area under investigation; ground-water investigations begun or finished during the year and those continued from the previous year; reports on ground water published or ortherwise released during the year; reports on ground water published in previous years; summary of water-level fluctuations during the year, comparison of water levels in the current year with the levels in previous years; summary table giving highest and lowest observed stages and dates on which the stages were observed; summary table giving net changes in individual wells or average net changes in groups of wells in the year; table of average water levels; data regarding consumption of ground water; detailed correlation of water-level fluctuations with precipitation, pumpage, or other influences; any other concise, pertinent discussion. *Well descriptions* — Well number used by Geological Survey, other number (if any), owner's name, location of well, use of well (if any), type of well, diameter of well, depth of well, description of measuring point, height of measuring point above or below sea level or other datum, description of bench marks, and other descriptive data that contributor deems advisable, such as statements regarding water-level fluctuations, influence of pumping, kind of pump or pumping equipment (if any), and operation of water-level recorder.

Wenzel's instructions have been followed, and the annual water-level reports (8) have become a valuable set of reference volumes for the ground-water hydrologists of the country. Improvements and additions are constantly being made both in the system of observation wells and in the reports for the purpose of making the records more usable and comprehensive. During recent years a monthly review (9) of the water resources in the United States and Canada has been issued, and this has included some information on about 160 of the key wells and occasional discussions of particularly unusual conditions in ground-water levels in various areas.

In several States, especially in California, extensive water-level programs are carried on entirely by State agencies.

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[7] Hydraulic Engineer, U. S. Geological Survey.

Measurement

In 1935, a committee [8] of geologists and engineers in the U.S. Geological Survey was appointed to prepare a manual of methods of measuring water levels in observation wells. In its report [10] this committee wrote, concerning the measurement of water levels in nonflowing wells : « Many methods of measuring the depth to water levels in wells have been devised, some of which are ingenious rather than useful. For most wells the method that is the most accurate and reliable, as also the most convenient, is to use a graduated steel tape with a weight at the end, the lower part of the tape being chalked and then let down below the water surface far enough to make a water mark on the tape. To the extent that reliability and accuracy are assured, it is desirable that the method used should, so far as practicable, be simple in operation, not time-consuming, and requiring only durable equipment that is readily portable and easily repaired or replaced. In general requirements are best met by the wetted-tape method. However, especially difficult conditions requiring other methods may be presented by wells with very deep water levels, wells having pumping equipment or other obstructions in them, and wells that are to be measured while they are being pumped or in which water may be dripping for any other reason. »

In addition to the wetted-tape method, the committee discussed measurement by the visible-ripple, sound, submerged-tube, air-line, and other methods. Concerning electrical methods, it was stated, « Many electrical instruments have been devised for measuring water levels in wells, They all operate on the principle that an electric circuit is closed when an electrode is lowered to the water surface in a well...

» Electrical instruments have advantages for measuring water levels under certain special conditions, as in ...wells in which water splashes from a pump pipe or from the wall of the well... For most purposes they are needlessly cumbersome and time-consuming and probably do not give as accurate results as the wetted-tape method. »

For flowing wells, the committee said : « In measuring the pressure head in flowing wells — that is, wells whose static levels are above the ground surface — it is necessary that the top of the well be closed so that shut-in pressure can be observed.

» ... In making periodic measurements of the shut-in pressure of a flowing well the time elapsed between closing the well and observing its pressure should be the same for all measurements, and the time should always be recorded.

» ... In flowing wells in which the water will rise only a few feet above the surface, a practical way of measuring the pressure head is to connect an ordinary garden hose tightly to the well and then to raise the free end of the hose until the flow of water just ceases. »

« ... Where static water levels are more than a few feet above the surface, it is generally necessary to use a device that indicates the pressure at a convenient point near the surface. The most common device for this purpose is the indicating pressure gage.

» ... A mercury manometer designed for field use gives somewhat greater accuracy than an indicating pressure gage and eliminates the need for periodic calibration. »

Concerning the frequency of measurements and automatic recorders, the committee reported : « The factors that should determine the frequency of water-level measurements are the types of fluctuation to be observed, the nature of the studies contemplated, and the available personnel...

» Automatic recorders afford the best means of observing... rapid and irregular water-level fluctuations...

» Obviously more frequent measurements and a greater use of automatic recorders are required for intensive investigations than for the determination of annual, secular, or long-term fluctuations...

» In general, automatic recorders may be divided into two groups — those for recording fluctuations of pressure head in wells where the water rises above the ground surface and may be shut in under artesian pressure, and those for recording fluctuations of water level where there is a free water surface below the top of the wells.

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[8] R. M. Leggete (Chairman of the committee, now a Consulting Geologist), L. K. Wenzel, (Secretary of the committee) R. C. Cady (deceased), S. W. Lohman (Senior Geologist), V. T. Stringfield (Senior Geologist), R. W. Sundstrom (Senior Hydraulic Engineer), and S. F. Turner (Senior Hydraulic Engineer).

» The type of recorder most generally used on non-flowing wells consists of a float that rises and falls with the surface of the water, a device for transferring the motion of the float to a recording device, and a time driving mechanism. There are many kinds of float-type water-level recorders, and although they differ in mechanical detail, they are all similar in basic principle. The recorder most commonly used by the United States Geological Survey in observation-well work consists of a float suspended from the instrument by a thin metal tape or cable, at the opposite end of which is a counterweight... The recording mechanism of these instruments consists of a cylindrical drum containing a paper chart that covers a period of a week, and this type is generally called a « weekly recorder ». In some makes the movement of the float produces movements of the chart drum and the recording pen or pencil is actuated by the time mechanism. In others the chart drum is moved by the time mechanism and the pen is actuated by the float. In still another type, known as the « continuous recorder », the chart consists of a sheet of paper several yards long, which provides for a whole year's record and is wound from one roll to another. This type requires attention only once a month or once in 6 weeks, to wind the clock and remove the used portion of the chart.

« Records of fluctuations of pressure head in a well that flows may be obtained with a recorder of the float type if the casing can be extended upward higher than the head of the water, leaving a free water surface. If this is not possible... and if the well can be closed by a valve or other means, records of the pressure may be obtained with a recording pressure gage. This gage consists of a device for transferring the fluctuations of pressure into angular movement of an arm carrying a pen, which makes the record, and a time mechanism. »

Although the committee on observation wells made its report nearly 13 years ago, almost no changes have been made since that time in the basic methods commonly used by the Geological Survey. Special methods of measuring water levels, such as by the use of echometers, have been developed, and research is being performed on methods of measuring water levels, recording them, and transmitting the records long distances by wire, all automatically by electricity, but these devices are costly and have not yet come into general use. Many cheaper devices such as chemical indicators (11) and high-low gages (12, 13) have proved to be quite useful in special instances, but for general application they are not as economical, reliable or convenient as the simple methods described by the committee in 1935.

Interpretation

It has been stated (14) that « Water levels in wells are measured primarily for three broad purposes : (a) To maintain a current inventory of available water supplies; (b) to make possible the prediction of supplies that will be available in the future; and (c) to make possible the prediction of stream flow during periods of no rainfall... Some of the technical reasons for measuring water levels are : (a) To define piezometric surfaces; (b) to record changes in storage; (c) to determine directions of flow; (d) to determine recharge from precipitation and surface water; (e) to determine natural discharge by evapo-transpiration and by flow into streams and other aquifers; (f) to determine effects of artificial withdrawal and recharge; and (g) to determine the hydraulic characteristics of aquifers. »

In 1936, in his introduction (15) to a symposium on fluctuations of ground-water levels, D.G. Thompson (9) wrote : « Studies of fluctuations of ground-water level are probably of most importance in connection with quantitative studies of the supply available for various purposes — whether for public or industrial supply of cities, for irrigation in agricultural regions, or merely for small domestic supplies in rural areas. In quantitative studies, observations of fluctuations of the water table or artesian head give useful information as to recharge of the formation, the discharge by seepage, spring flow, evaporation and transpiration, the permeability of the formation, the effects of pumping, and other factors involved in the determination of its safe yield. However, observations of such fluctuations are proving to be valuable in a variety of other studies. These include studies of the relation between rainfall and fullness of the ground-water reservoir as an aid to predicting

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[9] Deceased Senior Geologist, U. S. Geological Survey.

the low-water flow of streams. Although not much attention has yet been given to the opposite phase of the problem, the question may be raised as to whether fluctuations of the water table may also be used to show danger of flood conditions when heavy rains produce an unusual saturation of the ground. Other studies relate the effects of the artificial disturbance of the natural conditions affecting groundwater recharge and discharge, and hence the flow of surface streams. These include the building of dams and locks which may produce either advantageous or damaging rise of the water-table upstream from these structures and opposite effects downstream from them; the construction of drainage-ditches and canals, and works to change the courses of rivers, all of which may lower the water-table; the development of irrigation and sub-irrigation projects which may cause the rise of the water-table generally for useful results, but in many areas produce the opposite effect when water logging and spoilage by alkali occur; and efforts to conserve water by spreading the flood flow of streams that otherwise would be wasted. Fluctuations of water level have also yielded ... information of interest ... in regard to problems of the elasticity and compressibility of water-bearing formations, of transmission of earthquake waves, and of tidal phenomena. »

Water levels and artesian pressures in wells fluctuate in response to a great variety of natural and artificual phenomena, including artificial discharge and recharge. Meinzer has stated (16) « The water level in a well is sensitive to every force that acts upon the body of water with which the well communicates. Hence, in most wells the water level fluctuates almost constantly, often in a complicated manner. The curve produced by an automatic water-stage recorder over a well gives an accurate record of the resultant of the forces that act upon the ground water. »

Later he wrote (17)... « the way has been opened for making exact and critical studies of the dynamics of the ground water and of the properties of the ground-water reservoirs and conduits. However, as the fluctuations are seldom the effects of simple forces acting singly, careful selection of the observation wells and painstaking analysis of the records are necessary for evaluation of the effects of specific forces. » In that report he gave examples of « hydrographs showing fluctuations in water levels in wells produced by several different causes — precipitation, evaporation and transpiration, pumping, variations in atmospheric pressure, ocean tides, earth tides, earthquakes, and railroad trains. » Other causes of fluctuations that might be mentioned are variations in stream or pond stage, variations in temperature, infiltration of irrigation waters, variations in hydrostatic pressure in formations above or below that under consideration, artificial recharge, leaking wells, temporary freezing, changes in permeability or recharge conditions, and the like.

In the early part of the 20 century, when the importance of fluctuations of water levels was just beginning to be recognized, the interpretation of the fluctuations was largely accomplished by simple empirical methods. Later, more complicated graphical and statistical methods were put into use, and recently several workers have applied mathematics to the correlation of fluctuations with their causes.

Studies of rises and declines of the water table caused by variations in precipitation have been given particular attention for many years, for these are used in one way or another to measure recharge and natural discharge. In such studies it is generally necessary to know the specific yield of the water-bearing material through which the water table fluctuates. Wenzel (18), Theis (19), Williams and Lohman (20), Jacob (21), and many others have applied a wide variety of methods to the correlation of water-table fluctuations with precipitation.

Records of water-table fluctuations caused by evaporation and transpiration yield information concerning natural discharge, and lead to the determination of safe yields of shallow aquifers. White (22), Parker (23), and others have shown how this is done in practical investigations.

Until recently barometric, tidal, and loading effects were principally of only academic interest. It is now realized, however, that they yield a great deal of information concerning the elasticity or rigidity of aquifers, and the degree of confinement of the waters therein. For instance, Jacob (24) has presented a mathematical method of determining the storage coeffecient from barometric fluctuations.

In a number of areas it has been shown that geologic structures, not located by other means, are the causes of inconsistencies in water-level fluctuations in adjacent wells. The work of Poland and others (25) at Long Beach, California, is an outstanding example.

C.F. Merriam [10] (26) has been particularly successful in predicting low flow of the Susquehanna River by means of a water-level index based on fluctuations of water levels in shallow wells in the drainage basin of the river. Yonkers (27) and others also have made progress correlating water-table fluctuations with stream flow. This subject will be of major consideration in the enlarged Federal observation well network that is now being established.

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[10] Mechanical Engineer, Pennsylvania Water & Power Company.

Rorabaugh (28), Kazmann (29), Guyton, Stuart, and Maxey (30) and others have used fluctuations in the water table adjacent to streams to estimate possibilities of induced infiltration from the streams. This is another phase of ground-water work that has increased in importance, for modern construction practices are making it possible to withdraw very large supplies of water at low cost from wells and collectors in gravels that are recharged by adjacent streams.

In recent years the problem most often encountered in ground-water investigations has probably been that of predicting fluctuations of water levels that will be caused by future withdrawals of water. However, as stated by C.V. Theis [11] (31), « A knowledge of ... the essential factors that control the response of an aquifer to development of wells ... including the role of time, is necessary for the interpretation of existing records of water levels, and can yield the only method of predicting the effect of ground-water development in an area where records of long duration are lacking. »

Theis has had an important part in the determination of these factors. Excerpts from one of his principal discussions (31) are quoted as follows : « The essential factors controlling the action of an aquifer appear to be (a) the distance to, and character of, the recharge ; (b) the distance to the locality of natural discharge ; and (c) the character of the cone of depression in the given aquifer...

» All ground water of economic importance is in process of movement through a porous rock stratum from a place of intake or recharge to a place of disposal. Velocities of a few tens or a few hundreds of feet a year are probably those most commonly met with in aquifers not affected by wells. This movement has been going on through a part of geologic time. It is evident that on the average the rate of discharge from the aquifer during recent geologic time has been equal to the rate of input into it. Comparatively small changes in the quantity of water in the aquifer, with accompanying changes in water level, may occur as the result of temporary unbalance between discharge by natural processes and recharge, but such fluctuations balance each other over a complete season or climatic cycle. Under natural conditions, therefore, previous to development by wells, aquifers are in state of approximate dynamic equilibrium. Discharge by wells is thus a new discharge superimposed upon a previously stable system, and it must be balanced by an increase in the recharge of the aquifer, or by a decrease in the old natural discharge, or by loss of storage in the aquifer, or by a combination of these.

« Under Darcy's law there is only one way of reducing the flow in the areas of natural discharge or of increasing the flow in the areas of recharge. This is by changing the pressure gradient or the thickness of saturation of the aquifer in those areas, which in turn means changing the height to which water levels rise in wells throughout the area between the producing wells and the areas of natural recharge or discharge. This means a lowering of water level everywhere between the wells and the areas of natural discharge or recharge. In turn this means a reduction of storage in the aquifer and an abstraction of water from it.

» There are two fundamental properties of any aquifer which largely control the movement of water through it. The first is the ease with which it transmits the water... This characteristic of the aquifer as a whole is called the coefficient of transmissibility and is defined as the number of [U.S.] gallons of water that will pass in one day through a vertical strip of the aquifer one foot wide under a unit pressure gradient.

« The other important characteristics of the aquifer is the amount of water that will be released from storage when the head in the aquifer falls. This has been called the coefficient of storage, and is defined as the amount of water in cubic feet that will be released from storage in each vertical column of the aquifer having a base one foot square, when the water level falls one foot. For non-artesian aquifers the coefficient of storage is nearly identical with the specific yield of the material of the aquifer. For artesian aquifers the coefficient depends on the compressibility of the aquifer or of included or stratigraphically adjacent shaly beds and is much smaller.

» ... When a well is drawn upon... water levels are drawn down in the vicinity of the well. Some water is removed from the vicinity concurrently with this reduction in water levels, and a so-called cone of depression is formed. The shape of this cone is determined principally by the ease with which water flows through the aquifer — the coefficient of transmissibility — and by the coefficient of storage. ...With continued pumping the cone deepens and broadens. It is evident that the well is taking water out of storage in the vicinity and that as more and more water is removed by the well, the cone of depression affects more and more distant parts of the aquifer.

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[11] Senior Geologist, U. S. Geological Survey.

» The formula for the cone of depression in the ideal homogeneous and isotropic aquifer (of infinite lateral extent)... is :

$$s = \frac{114.6Q}{T}\int_{z}^{\infty} (e^{-u}/u)\, du$$

in which

s = drawdown at any point, in feet. [12]
Q = rate of discharge of the well, in [U.S.] gallons per minute. [12]
T = coefficient of transmissibility.
r = distance between pumped well and point of observation, in feet.
S = coefficient of storage. [12]
t = time the well has been discharging, in days.
$z = 1.87\, r^2S/Tt$. [12]
u = a dimensionless quantity varying between the limits given.

» The rate of pumping causes a proportional variation in the depth of the cone but does not affect its radius. The coefficient of storage, S, because of its relation to time, affects the rate of lateral spread of the cone, the rate of lateral growth being inversely proportional to its value. The coefficient of transmissibility affects both the radius of the cone and its depth, the radius for any given time increasing with increasing transmissibility, and the depth being inversely proportional to the transmissibility... The disturbance in the aquifer created by the discharge of the well may be likened to a wave : the amplitude depends on the strenght of the disturbance but the rate of propagation depends only on the medium in which the wave is formed. The reservoir from which the well takes water is almost as closely circumscribed by time as it would be by any material boundary, and until sufficient time has elapsed for the cone to reach the areas of natural discharge and rejected recharge a new equilibrium in the aquifer cannot be established.

» After the cone of depression reaches areas of rejected recharge or natural discharge, it is modified by the effects of adding water in the former or preventing it from escaping in the latter. If the rate of pumping does not exceed the amount of water added in the recharge area the cone will eventually reach equilibrium, at least practically speaking. The approximate effects that occur after the cone has reached the boundaries of the aquifer can be estimated by means of various mathematical analyses. The effects of discontinuous pumping can also be evaluated. »

Before Theis presented his « non-equilibrium » formula in 1935 (32), ground-water hydrologists had no method of really wide application for considering time in the mathematical interpretation of water-level fluctuations caused by pumping, although there were a number of « equilibrium » methods which do not take time into account. These have been described by Wenzel (33) and Meinzer (34), together with the use of the non-equilibrium method.

However, much information had been obtained on the effect of pumping on water levels as it varies with time ,and the long-term correlation of water levels and pumpage had been used to good advantage in making predictions of draw-downs and in determining safe yields. For instance, by empirical correlation of long term records of water levels and pumpage, after extraneous effects of precipitation, etc., were eliminated, D.G. Thompson and H.C. Barksdale [13] were able to determine the safe yield of the deep artesian aquifers at Atlantic City (35) and other places in New Jersey, and to predict successfully the declines which have since occurred as a result of increased pumping. W.N. White [14], at Houston, Texas, was successful in predicting by empirical means in 1937 (36) the effect of increasing the pumping rate in the area by 40 percent.

Since Theis' non-equilibrium formula was introduced, though, mathematical methods based on it have become more and more valuable, and predictions have not been so dependent on long-term correlations. Instead, the mathematical methods, which can be applied to the results of relatively short pumping tests, and which in some places are the only means of making predictions, have been used more and more extensively. These methods of making and evaluating pumping tests have been especially adaptable because of the thorough analytical study given them by C.E. Jacob [15] (37, 38, 39, 40).

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[12] Nomenclature changed from Theis' original paper.
[13] Senior Hydraulic Engineer, U. S. Geological Survey.
[14] Retired Principal Hydraulic Engineer, U. S. Geological Survey.

As W.F. Guyton and N.A. Rose [16] stated (41) in 1943 : « With the development of new methods, pumping tests have recently come into use. A pumping test and the use of its results are essentially a process of obtainig for the formation under investigation the equation of a water-level drawdown curve for a short period of time and extending the curve over a longer period by means of the equation. Predictions may thus be made of drawdowns caused by given rates of pumping. In making the predictions computed drawdowns must be adjusted for boundaries and changes in character of the formation not taken into account by the equation determined with the pumping test. »

Pumping tests have by now been made in numerous places all over the country, and wherever it has been possible to avaluate adequately the natural geologic conditions in the mathematical analysis, good results have generally been obtained. For instance, at Lufkin, Texas, it was possible to predict several years in advance, with errors of less than 5 percent, the drawdown of water levels of more than 125 feet resulting from doubling the pumping rate. At South Camp Hood, Texas, predictions a year in advance of drawdowns of 300 feet proved to be accurate within 3 percent.

As stated by Guyton and Rose (41) « ... pumping tests and their application have come into general use... At first they were considered too theoretical and impractical, but as they have been tried and developed more they have been found to be an exceedingly valuable addition to the methods used in ground-water investigations. They should, however, be used carefully and in connection with other methods wherever possible. Because a water-bearing formation is never as regular in character, thickness and extent as it is assumed to be by the formulas, its geology should be thoroughly investigated before the results or pumping tests are applied to determining drawdown. The reliability of the results depends very largely upon how well the geology agrees with the assumptions of the formulas.

» As the formulas are revised and now ones developed so that it becomes possible to take more of the geologic and hydrologic variations into account, the use and value of pumping tests and their application will become even greater than at present. »

The last statement is true also of all the other methods of measuring and interpreting water-level fluctuations. Recently a list of ground-water topics needing research was made in the Geological Survey, and the measurement or interpretation of water-level fluctuations is of major significance in 15 of 30 topics listed. Some of the items listed are : water-level measuring equipment; differences of water level between inside and outside of cased wells where pumped, the relation of ground-water fluctuations to the texture and structure of water-bearing formations, compressibility and elasticity of water-bearing materials, relation of ground-water to stream flow, infiltration and ground-water recharge, hydraulics of ground-water flow, and transpiration of ground water. A thoughtful review of such topics emphasizes to the ground-water hydrologist that, although much use is made today, as compared to a few years ago, of records of water-level fluctuations and although our methods of interpreting them habe been much improved, future uses are likely to be much greater and improvements in our methods to be rather widespread. The rapidly growing demand for more exact quantitative appraisal of our ground-water supplies cannot otherwise be met.

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6) VARIATIONS OF HYDRAULIC LEVEL AND OF ARTESIAN PRESSURE IN WELLS OF THE VENETO PLAIN; THEIR MEASUREMENT AND INTERPRETATION

by

Ing. Luigi VOLLO

The origin and alimentation of the Veneto region artesian layers

Geologic researches, bent to establish the Alps structure, succeeded to state that the fundamental base of the many mountains which surround on the North the Veneto region consists of schistous crystalline and impermeable rocks.

Consequently meteoric waters absorbed through infiltration or ground water streams (« carsici » phenomena) by the permeable rocks such as the calcareous (chalky), dolomitic and detrited ones, ecc., stay at the foot of every mountainous groups and flow down, supplying streams, torrents and rivers.

The hills surrounding the plain are formed by eruptive rocks or by Eocene chalks or by marls and clay, that is, by impermeable bulks.

Ultimately the mountainous docks of the Veneto Region are separated from the plain by tight barriers in which the valleys and the river-beds that flows into the Adriatic are situated.

The opinion still diffused that underground hydraulic layers of the Veneto plain find their origin and alimentation through the absorption of rains and snows, as it happens in the mountainous docks, is anyhow wrong.

But it is most certain that chalky mountainous reliefs where there are many splits, clefts and hollows known as Carsici phenomena, absorb a good quantity of meteoric waters; but carsica corrosion stops at the foot of every group where consequently interior idric circulation ends; for this reason, as we have already told, the waters absorbed in the surface, flow from the sides giving origin to surface or propped up sources face, flow from the sides giving origin to surface or propped up sources in the sweeping dregs that gather in the valley bottom.

Such dregs have a more or less big thickness; anyhow at a certain depth the valley sides reunite; for this reason Alpine and Prealpine rivers when flowing through the valley, have a surface run-off and a sub-river bed one, but their waters do not sink underground.

The rivers flowing from the mountainous valleys have wandered for thousands of years in the plain carrying in their flood a great quantity of differently composed

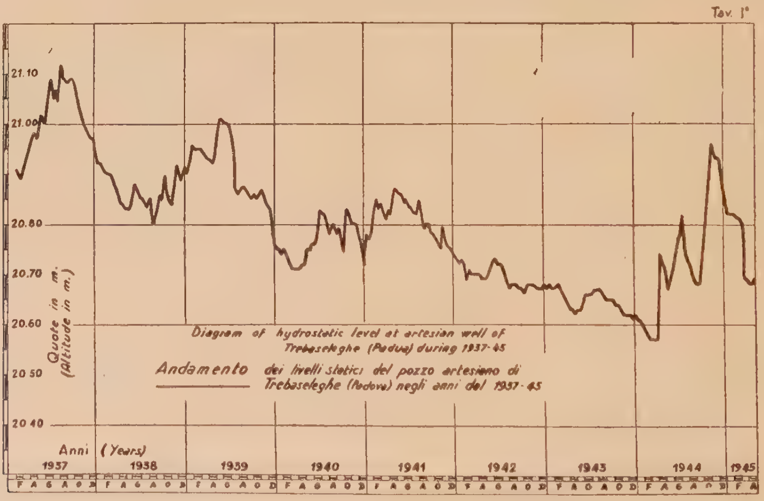

FIG. 1

sweepings, spreading in the shape of a fan, giving origin to big spoil mounds and covering the plain with those alluvial mantles that mix, near the coast zones, with sea deposit.

Every river along its mountainous flow enrich with its tributary streams contributions, and with the slow restitution of waters accumulated in the chalky bulks and in the sweeping heaps; when it goes out from the valley and enters the alluvial spoil mound, it looses a part of its run-offs through its gravelly bed; for this reason its discharges gradually diminish till they are completely annulled during low waters.

The spoil mounds are then imbued with the river waters that generated them by depositing for thousands of years every kind of materials which have placed alternatively permeable stratifications (gravels, sands) on impermeable ones (clays).

Such stratification are discontinuous, for this reason frequent argillous layers in the shape of lengthened lenses, alternate in the alluvial deposit. Such layers are irregular in thickness and inclination; sometimes twisted or elongated in digitations which take the strangest shapes; they alternate in their depth and often anastomose among them.

If the spoil mound is of a superficial kind, that is, it is placed on argillous soils then there is a crown of sources at its perimeter; but if such spoil mound lowers itself and spreads under impermeable covers, as it happens in the Veneto plain, hydraulic layers penetrate underground even to a noteworthy depth and extend towards the sea.

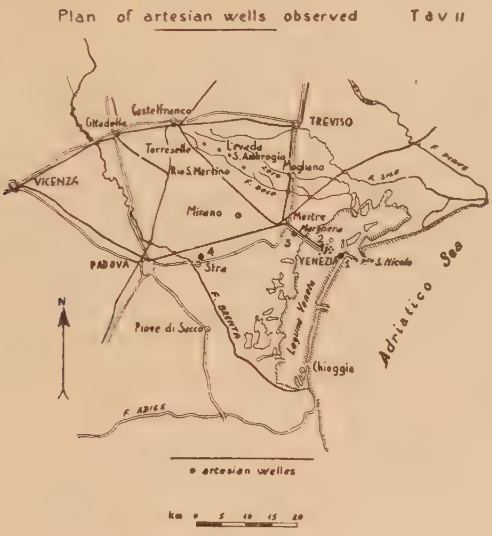

FIG. 2

The impermeable superficial septum forms the bottom on which the ground-water reservoir reposes and its bolder that extends towards the mount skims the Artesian waters that sink and remain imprisoned within the impermeable stratification.

Ascensional pressure of such layer appears heavier in the spot where the clayish lenses extend largely towards mountainous zones and reach higher levels.

Artesian waters movement or velocity is trifling; in an undisturbed regimen, it is some decimillimeter per prime minute. That happens only generally because, in our opinion, very frequently in the spot where solutions of continuity exist in the impermeable septa, the motion is considerably larger.

Anyhow, summarily, we can compare the Artesian layers with some big vessels of water under pressure having an overflow weir that goes to the ground-water reservoir and a sluice discharge to the sea; this sluice discharge is however half-closed on account of the high friction due to the porous division formed by small sands and the hydrostatic load accomplished by the sea in a variable quantity, according the tide oscillations.

Relation between tide to S.Nicolò-Lido and discharye of the wells during 18-1-1927

Tav. III

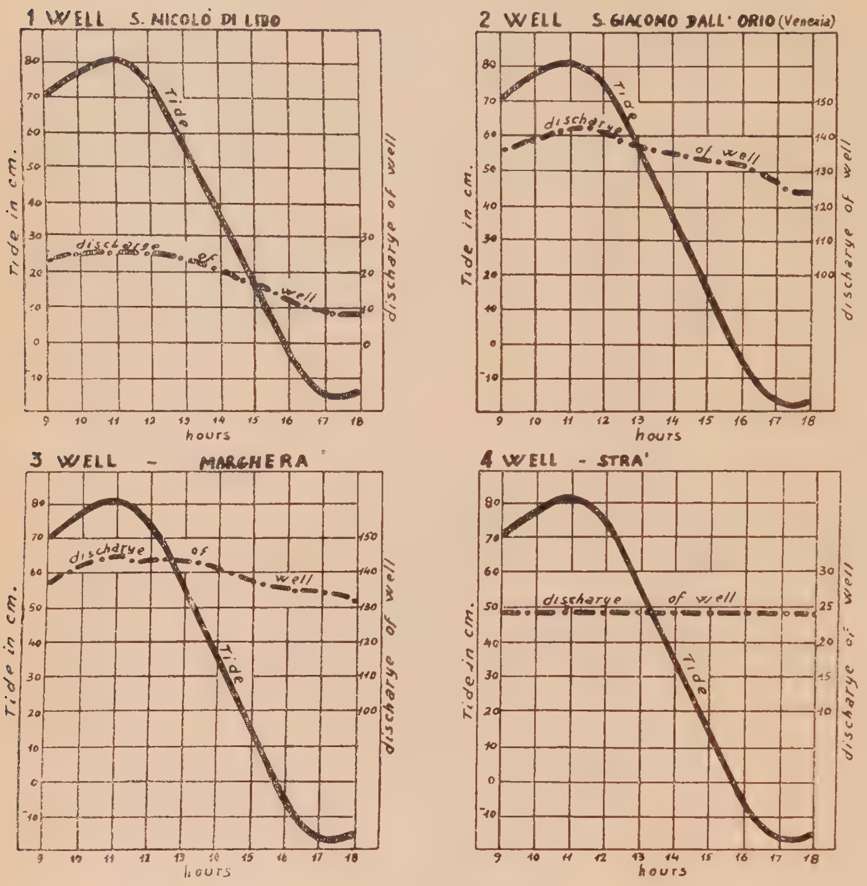

FIG. 3

This influence of the sea level excursion on the ascensional pressure of the artesian wells, was observed by us in 1926 on some bored fountains in the littoral drainages between the Piave and Tagliamento. On account of such ascertainment the Hydrografic Office of the Magistrato alle Acque disposed the testes of which we shall give a hint.

Level variations and ascensional pressure of the Artesian wells

The level oscillations of the wells opened on the Artesian layers and consequently those of ascensional pressure and discharges of the wells in distribution may be caused by different reasons whose effects are superposed and then it is very difficult to discriminate and analyse them.

The principal ones are the following :

1st — The barometrical pressure;
2nd — The hydrological conditions of the feeding docks of Artesian waters;
3rd — The tide excursions on the wells placed along the coast;

Relation between tide to S. Nicolò-Lido and discharge of the wells during 19-1-1927 Tav. IV

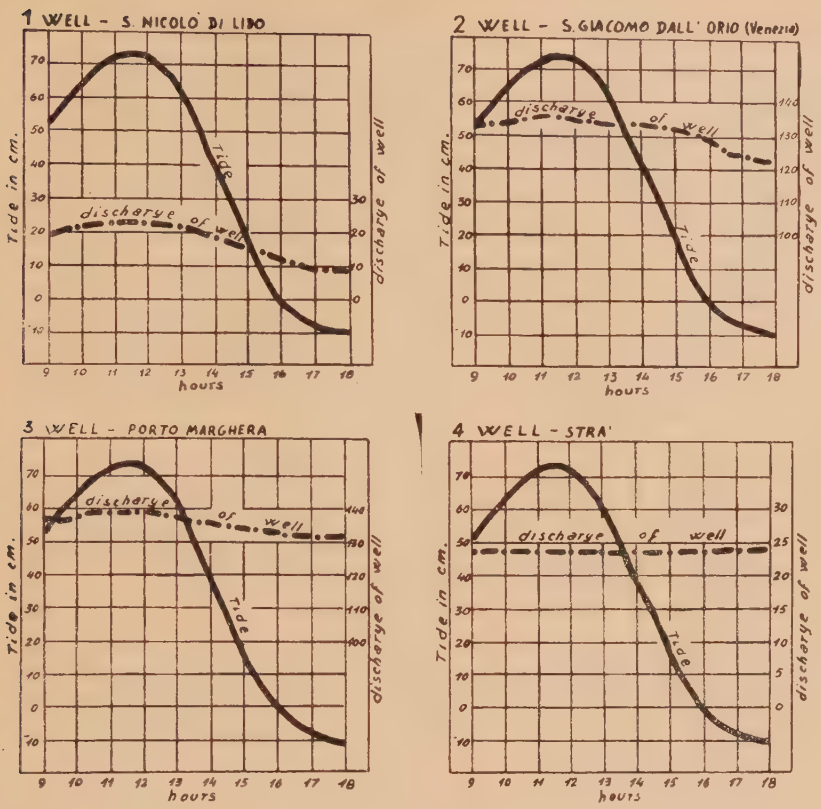

Fig. 4

4th — The free effluxion or suction drawings made in the same Artesian layer or in other layers in mutual connection with it.

Other hidden factors may have an influence on the variations of which we are discussing.

Barometrical pressure

Atmospherical pressure influence on the sources regimen and consequently on Artesian wells has been observed since 1881.

In that year in fact, Mr. Baldwin Latham gave a report on the observation and reliefs made by him of the sources regimen in the Caterham valley near Croydon and on deep wells, to the British Geological Association of York.

The experiments made in these last years by the « Magistrato alle Acque » Hydrografic Office of Veneto, have shown that while pressure variation have a noteworthy influence on the regimen of the natural sources and the ground-water reservoirs, such influence is however practically trifling on the ascensional discharge of Artesian wells.

Relations between tide to S. Nicolo-Lido and discharge of the wells during 17-18 February 1927 Tav. V

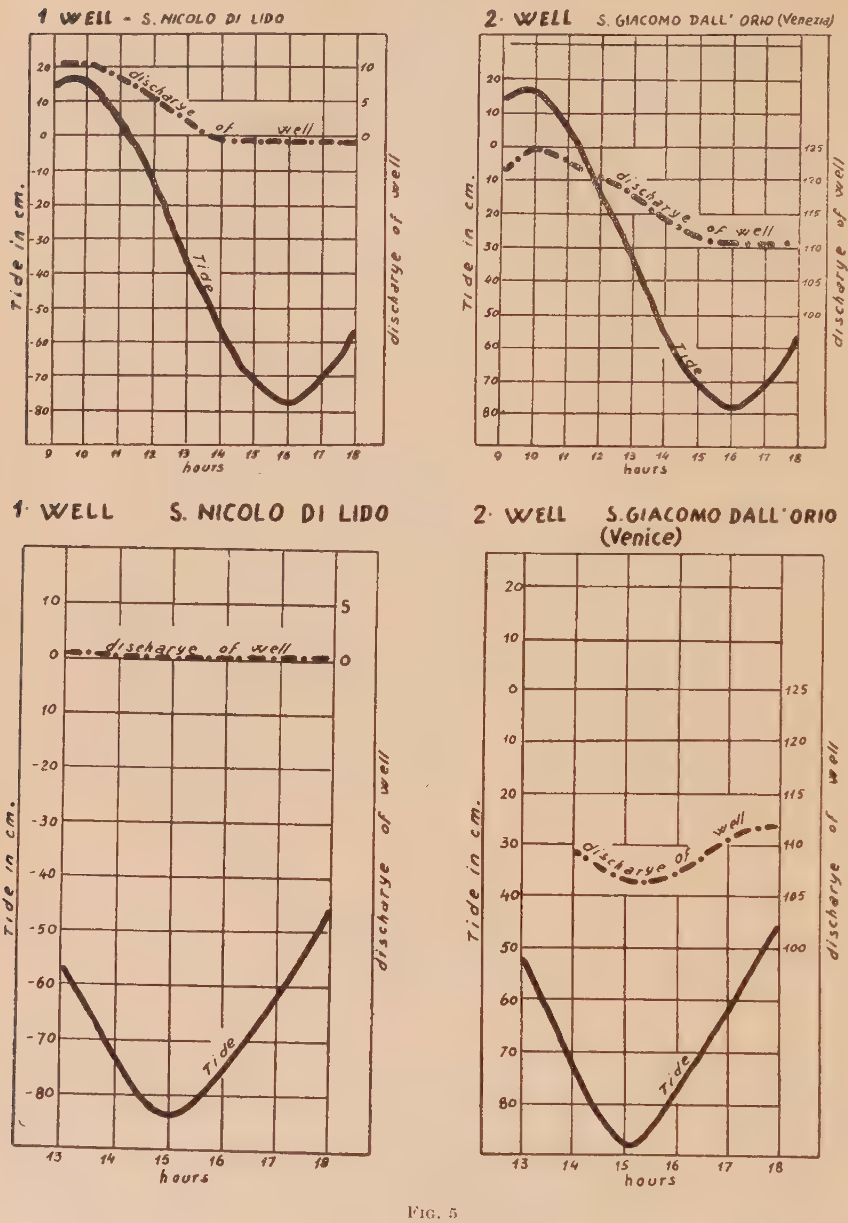

Fig. 5

Variations depending on hydrological conditions of Artesian layers feeding docks

What we have already stated shows without doubt that water level variations in the feeding dock consist of variations of pressure. This lowering is reflected on Artesian wells levels which generally can be considered hydraulic pressure heads of a closed conduit.

This representation of the Artesian wells with closed conduits is however only schematic and indicative and it does not permit to apply or to extend the hydraulic laws to the circulation and dynamism of the Artesian waters.

The place through which underground waters and mainly those in pressure flow, is very heterogenous; for this reason the filtration more than the circulation, is, in a different manner, made difficult by the numberless obstacles formed by the different composition and granulation of the permeable strata.

Losses of pressure are therefore noteworthy and very variable not only from layer to layer but, very often, from spot to spot of the same layer.

For this reason, while the hydric level variations of the feeding dock, or better, of the spoil mound are reflected almost immediately on some Artesian wells level, in some zones; such variation propagate extinguished and retarded, in others.

However, generally levels turn and consequently of Artesian wells ascensional pressure are interdependent and correlative to the hydrological turn (following to the meteorological events) of the layers feeding docks where wells are situated.

However seasonal and annual variations take place in the Artesian wells ascensional pressure. These variations are followed by diminutions or increasing of sources distribution discharge and natural effluxion and water-level reliefs in the wells utilized with raising installations.

In order to give an idea about the oscillations of a Veneto plain Artesian layer ascensional pressure,we have reproduced in the first table, here enclosed, the diagram of the static level variations of a bored well, by the Hydrografic Office at S. Ambrogio di Trebaseleghe (Padova) in the zone where many others Artesian wells alimenting the Acqueduct of Venice are placed.

The well was made 45,00 meters deep that is about 25 meters under the mean sea-level and the water layer in pressure is in contact of the filter for a thickness of 5 meters.

The mounting colomn is prolonged over the soil in such a way as to keep the well at the hydraulic pressure head whose oscillations are registered by a depth recorder.

We must notice that according to the needs of the acqueduct working some distributory wells are opened and enclosed in fixed periods (summer-automn) and that the effect of such manœuvre is felt by the testwell by lowerings or raisings of level, of 1 — 3 centimeters.

However, the examination of the graphic, as for what concerns the natural process of the well levels, must be effectuated by a general vision of the curve turn, prescinding from the detailed examination of the secondary little ondulations that are not always due to hydrological variations.

As we can see by the diagram whose ordinates state the values of the decade averages registered in the period 1938-1945, the highest altitude (21.121 meters on medium sea-level) was reached in August 1937, which was a very rainy year; the lowest altitude of the period before mentioned (20.57 meters) was registered in March 1944, after a long dry period interrupted by heavy rains in April of the same year, in June and, with greater intensity, in October and November.

The highest level excursion was registered in a period of seven years which was characterized by a very rainy year and by four dry ones; such excursion (of 0.55 meters) is not noteworthy and it shows that the underground water-reservoir which feeds the layer has a big level volume.

On the other hand the rapid increase which took place on the year 1944, owing to the rains that though they were not exceptional, determined river's high waters. It shows the sure influence of the hydrological factor on the ascensional pressure of the layer.

Tide influence on the pressure variations of the Artesian wells

In order to establish if the tide influence on the Artesian wells is felt even at a certain distance from the coast, the Hydrografic Office, in 1927, has made some

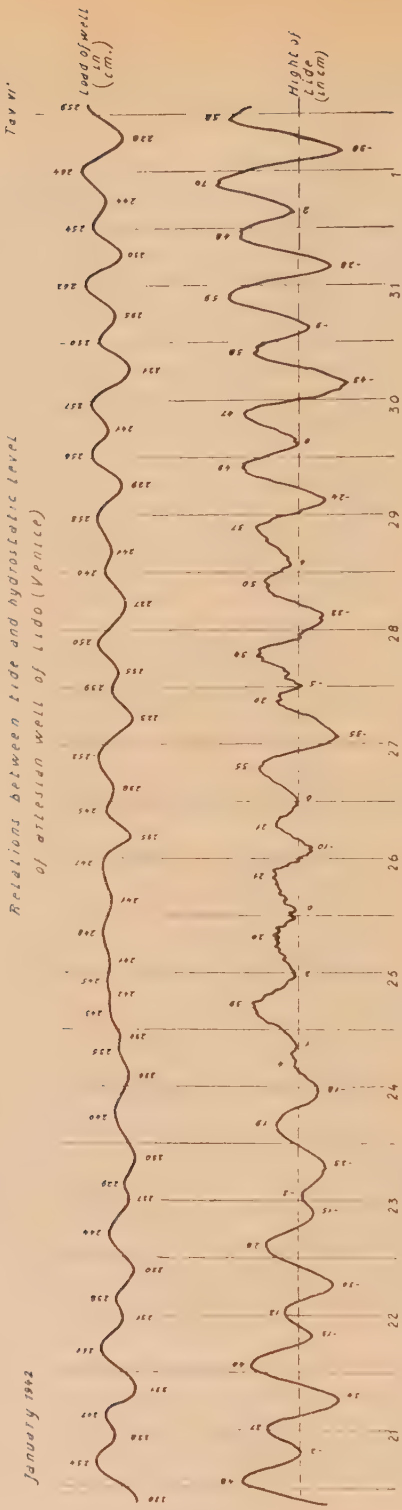
Tav VI
Relations between tide and hydrostatic level of artesian well of Lido (Venice)
January 1942
Load of well in (cm.)
Hight of tide (in cm)
21
22
23
24
25
26
27
28
29
30
31
1

FIG. 6

reliefs on four Artesian wells chosen on the straight way Venezia-Padova (see table 2nd) and placed exactly as follows :

1st) S. Nicolò di Lido, 600 m. for from the sea; a well 110.00 m. deep;

2nd) Venezia, Campo S. Giacomo dell'Orio, 5 Km. from the sea; well 180 m. deep;

3rd) Mestre, Porto Marghera, 11 Km. far from the sea; a well 200 m. deep;

4th) Stra, Ville Nazionale, 25 Km. far in straight line from the sea; a well 170 m. deep.

Contemporary and continuous discharge measurement during high and ebb tide phases were accomplished in such wells formed of metallic pipes of a diameter osciliating about 10 centimeters. They are supplied with a filter in that part which is immerged in the watery layer and naturally flowing at about a meter over the soil.

Such reliefs were repeated in three periods and the discharge values noticed in the four wells, in each period, were reproduced in the graphic with the tide turns registered by the depth recorder of S. Nicolô di Lido.

As we see, in the diagrams (3rd and 4th tables) with a tide excursion of about 1.00 meter (from 0.82 above to 0,18 below the medium sea level) the discharge of Lido well has changed from a minimum of 8 liters to a maximum of 24 liters per minute with an increase of 300 % on the minimum; the discharge of Venezia well from a minimum of 124 liters per minute has raised to a maximum of 143 liters with an increase of 15 %; the discharge of Marghera well has increased from 133 liters to 145 with an increase of 9%; while in the well of Stra there have been no variations.

Having repeated the reliefs during the ebb tide of February 17th and 18th in 1947 (see 5th table), we have noticed that when the tide height lowers of 0.50 m. below the medium sea level, the well of S. Nicolò di Lido stops its flowing, while in the well of Venezia the pressure lowers to 107 liters for minute corresponding to a tide height of 0.89 m. below the medium level.

Then we deduced that the tide influence on the Artesian layers, which was very remarkable along the coast, diminished towards dry-land till it finished completely when at 15 Kilometers far from the coasts.

All that happens for deep layers, along the straight way Venezia-Padova that is from East to West.

But successively, having tested variations of discharge of other wells in the territory of Mogliano Veneto that is at the North of Venezia (see 2nd table) and at a distance of about 23 Km. from the sea we have noticed that the Artesian layer which is 45.00 meters underground (about 35 meters below the medium sea level) and which feeds numerous fountains of that zone, feels in a noteworthy manner the tide oscillation. In fact a tide excursion of about 40 centimetres has provoked an increase of pressure of 10 : 15 centimeters with a delay of many hours.

Then the repercussion of the tide oscillations is felt towards North and North-East at more than 25 Km. far from the coast.

Some observation data have been gathered from 1940 to 1943.

The level oscillations of the well in a static regimen were registered at a depth of 105.00 meters in the well lately bored, near the shore of Lido (Palazzo del Cinema).

The tide diagrams and the level variations of the Artesian well are reproduced in the same scale in the 6th - 7th - 8th enclosed tables.

If we examine superficially the graphics we notice that the two diagrams curves go straight on parallel with a delay of about three hours in the oscilliatons of the well level with respect the to tide excursions.

The ratio among the amplitudes of the excursions seems kept included between a half and a third.

However, if we analyse such excursion we notice that the delay is variable and that ratios between the well level variations and those of the tide always change and oscillate from a minimum of 0.11 (1) to a maximim of 0.48 (2). We have observed such extremes values in the same period of squaring.

Normally ratio values oscillate between 0.30 and 0.40; for that reason very often level variations in the wells are in measure of 1/3 of the tide excursion.

Other tests were accomplished in the Artesian wells in Venezia and concerning the layer situated about 110 meters underground.

We have had the most evident results in the relief fulfilled on December 27th 1944 when we had a tide excursion of 110 cm. to which a pressure variations of 35, 32, and 36 cm. have corresponded in the wells of S. Maria Formosa, Campo degli Arditi and Campo SS. Givanni and Paolo.

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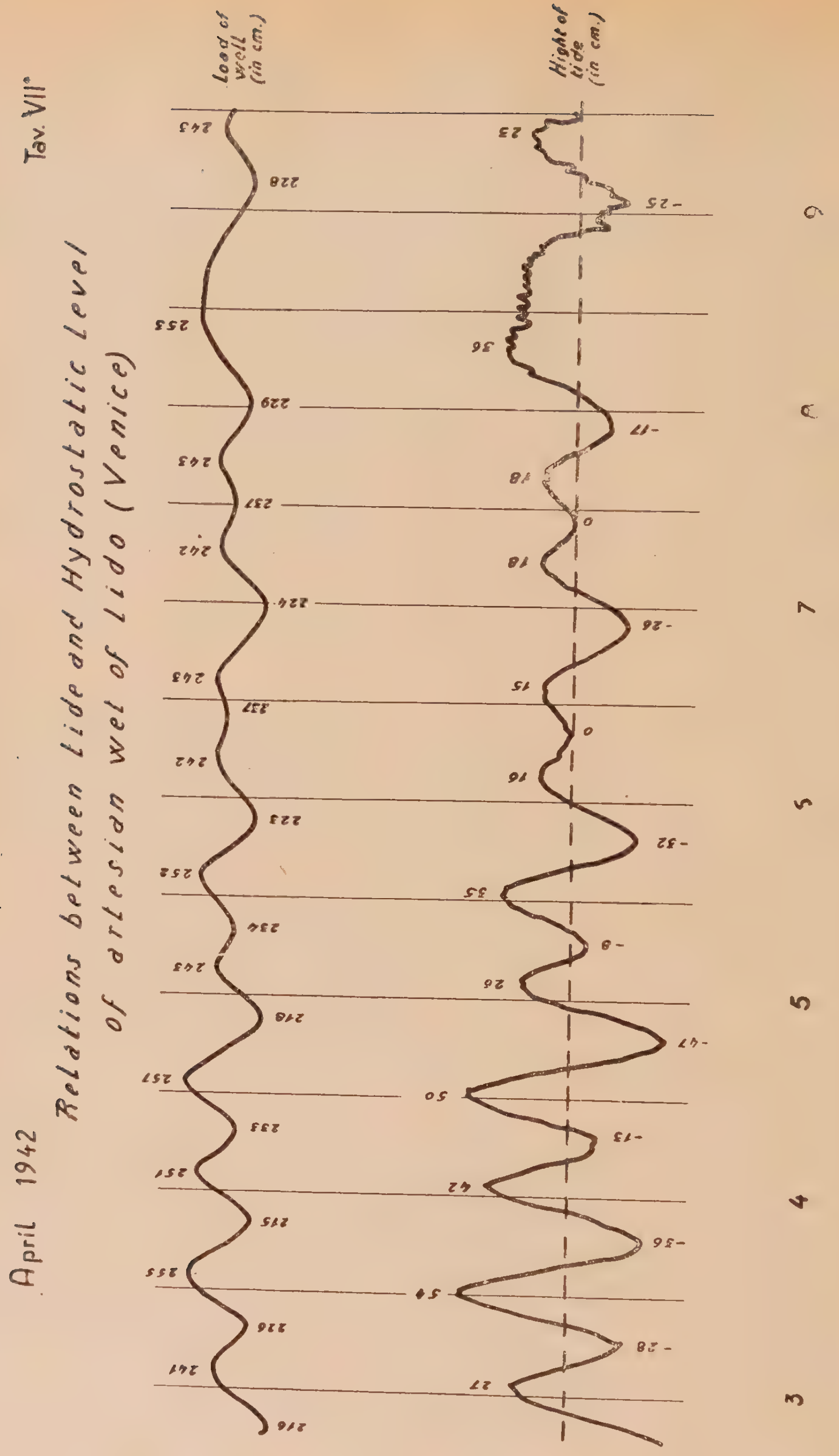
(1) on January 25th 1942 we observed a lowering of 4 cm. in the well and a sea level lowering of 37 cm.

(2) between the 23rd and 24th of January 1942 to a raising of 20 cm. in the well corresponded a raising of tide of 42 cm.

Tav. VII°
April 1942
Relations between tide and Hydrostatic Level of artesian wel of Lido (Venice)
Load of well (in cm.)
Hight of tide (in cm.)
216
241
216
255
215
251
233
257
218
243
234
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223
242
237
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243
27
-28
54
-36
42
-13
50
-47
26
-8
35
-32
16
0
15
-26
18
0
18
-17
36
-25
23
3
4
5
6
7
8
9

FIG. 7

We must notice that the layer in communication with the wells of S. Nicolò di Lido of the shore of Lido (Palazzo Cinema) and of Venezia city is found at the depth of about 110 meters under the medium sea level while the bottom of the high Adriatico in front of Venezia reaches its greatest depth at 35.00 meters.

We have seen that the well of Venezia and also those of Marghera, which catch the waters of a layer 180.00 meters deep, feel the tide influence.

If the deep Artesian water should flow into the sea they should raise by filtration, through a sandy deposit, necessarily compact, in the first case, eighty meters thick and in the second one, of about 150 meters thick.

In the case that such filtration may happen, which is rather incertain the the frictions would certainly neutralize the scarce hydrostatic load constituted by the tide excursions which in the cases observed and also in a period of dryness normally does not exceed a meter height.

On the contrary, we have seen that small sea level excursions of 10 — 15 centimeters produce sure level variations of 2 — 3 centimeters into the wells.

Then it is evident that the deep layers are not in communication with the sea. According to our opinion the explanation of the phenomenon is shown by the fact that the superposed Artesian layers, though separated by impermeable strata, are in mutual connection through the continuity solutions of the same strata.

The anastomosis process of some watery layers in pressure has been ascertained more than once by us and it has demonstrated unwarranted the hypotesis maintained by clever schemers that the exploitation of a layer could not reflect on the others standing above.

Six hydraulic layers in pressure have been found out and examined closely in the underground of Venezia.

— The first 15 ÷ 20 meters underground with a very limitated ascensional pressure, so that the waters do not raise.

— The second 55 ÷ 65 meters underground with an ascensional pressure sufficient to make the waters raise to the soil level which is situated one meter above the medium sea level.

The third 105 ÷ 110 meters underground with a moderate ascensional pressure (about 0.80 meter above the soil).

— The fourth 150 ÷ 160 meters underground also with a moderate ascensional pressure.

The fifth 200 ÷ 220 meters underground also with a moderate ascensional pressure.

— The sixth 270 ÷ 280 meters underground with a rather important acensional pressure.

We think that only the first layer is in communication with the open sea, at a short distance from the coast where the same depth's lines oscillates between 10 and 20 meters and that such communication shows itself by stillicide through thin sandy strata.

Below Artesian layers in communication with the above layer would feel the hydrostatic load prevoked by the tide oscillations.

We have related the facts and alluded to an hypothesis but we leave the field open to whom, will be able to deduce other illations from the phenomenon observed and stated by us.

The influence of the function of the Artesian wells on the regimen of watery underground layers

Generally one think that the opening and the closing of one or more Artesian wells can provoke alterations of ascensional pressure of the only concerned layer of the same wells.

Instead, by what we could ascertain, we have seen, in many cases that such alteration reflect in more or less considerable measure on all the above watery layers and in some specified cases also on the ground water's layers of the high plain.

The forced limitation of space does not consent to coordinate and state the material of observation collected in the inquiries fulfilled for the study of each problem concerning the exploitation of Artesian waters.

We shall only give a short hint about two sure facts.

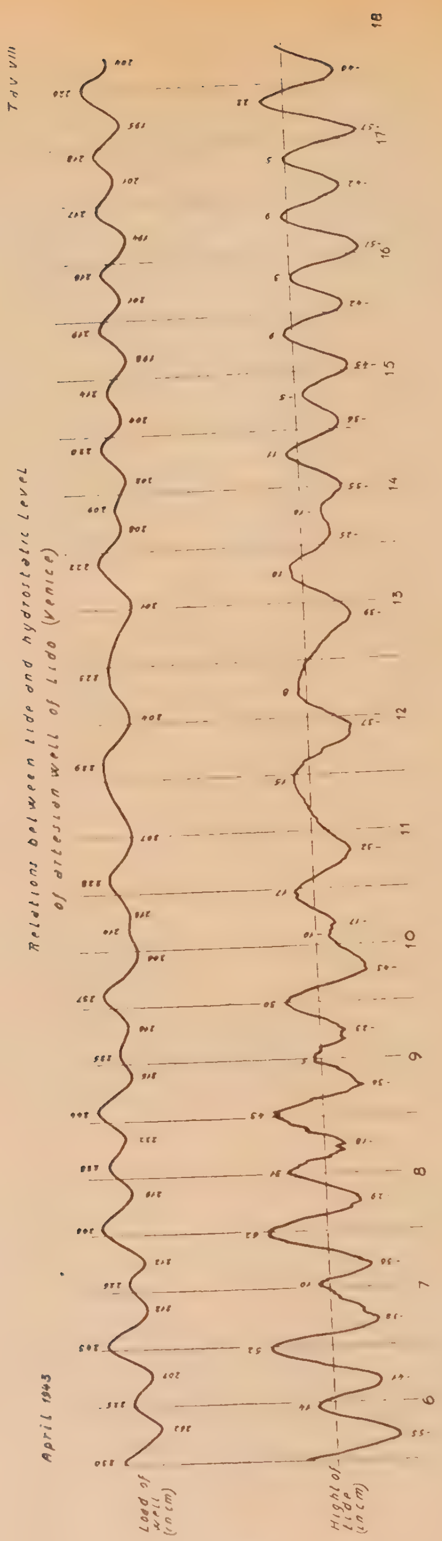
Tav VIII
Relations between tide and hydrostatic level of artesian well of Lido (Venice)
April 1943
Load of well (in cm)
Hight of tide (in cm)
6
7
8
9
10
11
12
13
14
15
16
17
18

FIG. 8

Exhaustion of ascensional pressure in the Artesian wells of Venezia

About thirty Artesian wells for hydraulic supply were bored in Venezia from 1848 to 1870.

At the beginning of our century the dicharges flowing out from such wells began to diminish, on account of the filters and pipes wearing out; for this reason the boring of the new wells, near those already existing or in new places, was made from 1913 to 1915.

Generally the borings were made 110 meters below the soil level. The penetration depth was limitated to 50 ÷ 60 meters for some wells.

The discharges flowing out, about 1 meter above the soil, were abundant (they oscillated about 90 ÷ 120 liters per minute) for diameters of 100 mm.

After the war 1915-1918 the industrial Harbour of Marghera reached a remarkable development and it was necessary, on account of the industries need, the perforation of Artesian wells till to concerned layers 200 ÷ 250 meters deep.

However such wells did not distribute freely the very large quantity of waters necessary for refrigeration and industrial workings. For this reason there were put special submerged pumps for drawing and hydraulic raising.

Very large volumes of water were absorbed underground causing level lowering in the wells that lowered till 20 meters below the plain country level.

At the same time, other borings of wells, concerning the same deep layers, were made in Venezia and in the isles for other industrial uses, from 1920 to 1934.

The effect of such drawings was also felt in the already existing wells bored 60 meters and 110 meters underground which ceased to distribute and were considered exhausted.

However it was only due to lowering of the ascensional pressure which has caused a lowering of the static level below the soil surface.

In fact, we advised to place some common piston pumps on such wells,thought exhausted, and so it was possible to utilize them again through drawing.

The lowering of ascensional pressure did not happen only to Venezia wells, but to all the wells bored on dry land in the territory East and West of Mestre. in which effluxion ceased.

Owing to air bombardment of Porto Marghera and to war vicissitudes, the industrial activity of the same harbour was paralysed in the years 1944-1945. So that drawing through suction ceased completely.

Notwithstanding hydrological exhaustion due to the dryness period occurred between 1940 and 1945 and lasted till 1947, which has caused a lowering of pressure of all underground layers. We noticed in consequence of the limitation of deep layers utilization at Porto Marghera a progressive low increase of levels of all Venezia Artesian wells enclosed those concerning the layer existing at 55 ÷ 60 meters. This is a sure demonstration of the regimen interdependance of the many layers.

Artesian wells of Lonigo in the province of Vicenza

The exploitation of Artesian waters was made since 1870 in the Almisano hamlet and precisely in a zone having a surface of about 3000 hectares belonging to the large alluvial spoil mound formed with the secular dejection of the rivers Agno and Chiampo, in which more than 400 wells were bored and in distribution in 1936. These wells were formed by metallic pipes of diameter changeable between 30 and 100 millimeters, infixed to a deepness of 24 ÷ 32 meters in which the static level raised from 3 to 7 meters above the country plain in connection to the altitude of the ground.

The exploitation of the Artesian layer below the one utilized and which was placed at deepness of 60 meters, was decided in consequence of the initiative of building an interprovincial acqueduct fed by the Artesian waters of Almisano and having been noted the gradual lowering of pressure of the existing wells, as a consequence of new borings.

The presence of spongy rock which is thought to belong to the Eocenic period and which formed the very ancient bottom of the valley was found by underground exploration made by a test well at the depth of about 80 meters, that is 50 meters below the medium sea level.

The schemers'idea about the existence of a layer independent from the above one and supplied with a greater oscensional pressure has been encouraged by the fact that static pressure would raise till 10 meters above the ground in the test well.

The Hydrographic Office, having to give an opinion about that, has made some tests in order to ascertain the source of the waters in pressure and the harmlessness of the planned installations on the existing utilizations.

Researches and tests were preceded by levellings to determine the altitudes of the many existing wells and the test-well.

The layer fed by the waters flowing from the Agno and that one fed from the Chiampo was found out by chemical analyses of the waters and temperature measurement.

A series of wells bored in the layer 30 meters deep, the test-well and other two wells bored in the layer at 60 meters depth were placed on a static level. From that we deduced that the profile of the hydraulic pressure heads is regular enough and shows that the two layers get the same ascensional pressure.

Then some wells of the first layer were closed and the discharges and the static levels of contiguous wells bored in same layer and in the one below, were kept under observation for two days. Such manœuvre was repeated four times from the sixth to the 11th. of January and from the 23th to the 28th of February in 1928. So that we deduced that level variations and lowering and raising of discharge following to the stopping manœuvre and re-establishment of distributions happened in a different measure, but in all the wells kept under observation. Which shows that the two layers are interdependent.

Measurement of ascensional pressure variations of the wells

The best system in order to follow the ascensional pressure variations of the waters in an Artesian well is to lengthen the mounting column in such a way as to put the well on a static level.

If we apply to its summit a depth recorder with a little reduction ratio (from 1/2 to 1/5), we can see with great precision the fluctuation of ascensional pressure.

In the cases in which such pressure is so high and such to raise the static level till to many meters above the ground, it is easier to measure instantaneous pressure by putting the well in effluxion and applying an usual manometer on the outlet.

The measurement of the discharges flowing at different heights from the ground, makes also possible to control the pressure variations that may intervene some time after.

The use of a recording gauge to follow the pressure variations through the time, is advisable only when it is possible to shut the apparatus in a room where the temperature is nearly always constant unless we use a mercury autocompensated manometer.

Venezia, February 26th 1948.

(Ing. Luigi Vollo)
Director.

COMMISSION DES EAUX SOUTERRAINES

Rapports relatifs à la troisième question :

SOURCES THERMALES ET COMPOSITION CHIMIQUE

1) LA THERMIQUE DES EAUX SOUTERRAINES D'ORIGINE PROFONDE

par

Henri SCHOELLER
Maître de Conférences de Géologie
à la Faculté des Sciences de Bordeaux

Dans un précédent travail (1), j'ai examiné la thermique des eaux des sources d'origine peu profonde, et circulant dans les terrains, sous des inclinaisons voisines de l'horizontale. J'étudierai ici le cas des eaux d'origine profonde et se mouvant verticalement ou presque verticalement dans l'écorce terrestre. Comme dans la première étude, je me propose d'abord de rechercher quelle est la température de l'eau aux différents points de son circuit, que cette eau soit ascendante ou descendante. Ensuite j'essaierai d'examiner une relation importante existant entre le flux de chaleur de la terre et la température de l'eau à la source, et dont on doit nécessairement tenir compte dans l'appréciation de l'origine de l'eau.

Je ne puis évidemment prendre que des cas théoriques. Certes ils ne permettront pas de définir dans un système hydraulique naturel, d'une manière précise, la température à tel ou tel point de circuit. Mais ils indiqueront la marche du phénomène. Et c'est en cela qu'ils pourront être intéressants.

Cas des chenaux subverticaux ou verticaux, ascendants.

Supposons que nous ayons un chenal simple, rectiligne et cylindrique. Une telle régularité, certes, n'existe pas dans la nature. Supposons que notre chenal soit situé dans un terrain de température donnée, sans flux de chaleur, ou recevant un flux de chaleur de toute part dirigé perpendiculairement à la conduite. Soit ρ le rayon du chenal dans lequel l'eau circule à la température θ en un point donné. A la distance R_1 de l'axe du tube, se trouve l'isotherme θ_1 dans un plan normal à l'axe du tube. Soit P_0 le flux de chaleur passant à travers l'unité de longueur du tube quand l'état permanent est atteint. Admettons que θ_0 soit à la fois la température de l'eau et celle de la paroi de la roche au contact de l'eau, ce qui est admissible, puisque l'on peut négliger les phénomènes de convection, en raison de la faible conductivité du terrain et de la petitesse du flux de chaleur. On a alors :

$$P_0 = 2\Pi RK\, d\theta/dR$$

R étant le rayon d'un isotherme ; $d\theta/dR$, le gradient thermique et K la conductivité calorifique du terrain.

Par conséquent l'on a avec les conditions aux limites : $R = \rho$ pour $\theta = \theta$ et $R = R_1$ pour $\theta = \theta_1$:

$$P_0 = \frac{2\pi K(\theta_1 - \theta)}{\log e\ R_1/\rho}$$

et il pénètrera, le long de la longueur dl de la conduite :

$$P = \frac{2\pi K dl(\theta_1 - \theta)}{\log e\ R_1/\rho}$$

L'eau reste en contact de dl pendant le temps $dt = dl/V$, V étant la vitesse de l'eau. Elle recevra, donc, sur cette longueur dl, c étant la chaleur spécifique de l'eau :

$$c\,q = 2\pi K\ \frac{\theta_1 - \theta}{\log\ e\ R_1/\rho}\ \frac{dl^2}{V}\ \text{calories}$$

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(1) H. Schoeller. La température des eaux souterraines vadoses (= d'origine météorique). Mémoire de 168 p., 69 fig., Archives, n° 265a et 165b du Centre de documentation du Centre national de la Recherche scientifique, Paris 1947.
et H. Schoeller, Etude sur la température des eaux souterraines. Proc. verb. de la Sté des Sciences physiques et naturelles de Bordeaux, Séances du 10 et 24 avril 1945.

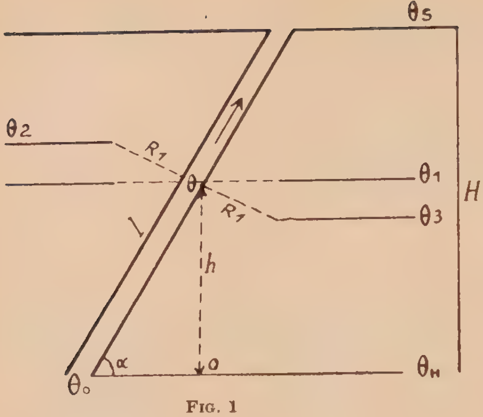

FIG. 1

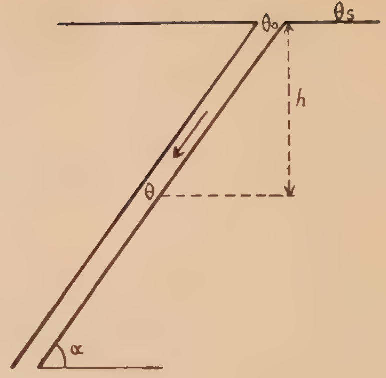

FIG. 2

la section du chenal étant $\pi\rho^2$, la température de l'eau augmentera donc de :

$$d\theta = \frac{cq}{\pi\rho^2 dl}$$

et l'on aura :

$$\frac{d\theta}{dl} = \frac{2K(\theta_1 - \theta)}{cV\ \rho^2 \log\ e\, R_1/\rho} = N\ (\theta_1 - \theta)$$

en posant

$$N = \frac{2K}{c\,V\ \rho^2 \log e\, R_1/\rho}$$

En intégrant, avec conditions aux limites $\theta = \theta_o$, pour $l = 0$, on a :

$$\theta = \theta_1 - \ (\theta_1 - \theta)\, e^{-Nl}$$

Supposons maintenant que nous ayons un chenal vertical ou subvertical, plongé dans des couches de plus en plus froides vers le haut, c'est-à-dire traversées par un flux vertical de chaleur et dirigé de haut en bas, de gradient thermique G (le gradient thermique terrestre étant environ de 1/35000, les unités étant prises dans le système C.G.S.). Dans le cas envisagé ici, on admet que le flux de chaleur arrive perpendiculairement à la conduite, de régions à température θ_2 et θ_3. Nous supposerons que cela revient au même que si le flux de chaleur arrivait de l'isotherme θ_1 intermédiaire, correspondant à l'ordonnée h, les deux effets se compensant plus ou moins, supposition d'autant mieux permise que la couche est plus voisine de la verticale.

L'équation (1) nous montre que sur une longueur infiniment petite $dl = dh/\sin\alpha$, la température de l'eau augmentera de

$$d\theta = N\ (\theta_1 - \theta)\ dh/\sin\alpha \tag{3}$$

Or si nous prenons comme origine, un point situé à la profondeur H sous le sol, point où la température est θ_H, la température θ_1 du terrain à l'ordonnée h sera : $\theta_1 = \theta_H - hG$

L'équation précédente devient donc :

$$\frac{d\theta}{dh} = -\frac{N\theta}{\sin\alpha} - \frac{NGh}{\sin\alpha} + \frac{N\theta_H}{\sin\alpha} \tag{4}$$

En posant : $a = N/\sin\alpha$ $\quad b = NG/\sin\alpha$ $\quad c = N\theta_H/\sin\ \alpha$

$$\frac{d\theta}{dh} = -a\theta - b\,h + c$$

qui intégrée donne :

$$\theta = \frac{-a\,b\,h + c\,a + b}{a^2} + \mu e^{-ah}$$

Si pour $h = 0$, $\theta = \theta_0$, il vient :

$$\theta = \theta_0 \, e^{-ah} + \frac{c\,a + b}{a^2}\,(1 - e^{-ah}) - \frac{a\,b}{a^2}\,h$$

d'où pour $h \neq 0$

$$(5) \qquad \theta = \theta_H - G\,h + (\theta - \theta_H)\,e^{-\frac{Nh}{\sin\alpha}} + \left(\frac{G}{N}\sin\alpha\right)\left(I - e^{-\frac{Nh}{\sin\alpha}}\right)$$

dans laquelle :

θ est la température de l'eau, à la hauteur h au-dessus de l'origine,
θ_H est la température du sol à la profondeur H,
$\theta_H - G\,h$ est la température du sol à la hauteur h,
θ_0 est la température de l'eau à l'origine à la profondeur H
G est la gradient géothermique,
α est l'inclinaison du chenal sur l'horizontale,

$$(6) \qquad N = \frac{2K}{c\,V\,\rho^2 \log e\,R_1/\rho}$$

comme précédemment (2), dans lequel :

K = le coefficient de conductibilité calorifique du terrain, en moyenne 0.005 en unités C.G.S.) c = la chaleur spécifique de l'eau et V = la vitesse de l'eau, ρ = le rayon du chenal cylindrique,
R_1 = le rayon du faisceau de flux de chaleur dont les isothermes sont modifiés R_1 étant mesuré perpendiculairement à l'axe du chenal.

Cette équation suppose que R_1 reste le même du haut en bas chenal ; ce qui évidemment n'est pas. C'est ce qui rend la formule approximative. Néanmoins comme R_1 rentre dans la formule sous forme de logarithme, l'approximation, de ce fait, se trouve fortement diminuée.

Si l'on appelle θ_S la température de la zone neutre du sol, on voit que pour avoir la température de l'eau de la source, il suffit de remplacer $\theta_H - G\,h$ par θ_S et h par H.

Si à l'origine, c'est-à-dire à la profondeur H, la température de l'eau est égale à celle du terrain, $\theta_0 = \theta_H$, la formule (5) se simplifie

$$(7) \qquad \theta = \theta_H - G\,h + \left(\frac{G}{N}\sin\alpha\right)\left(I - e^{-\frac{Nh}{\sin\alpha}}\right)$$

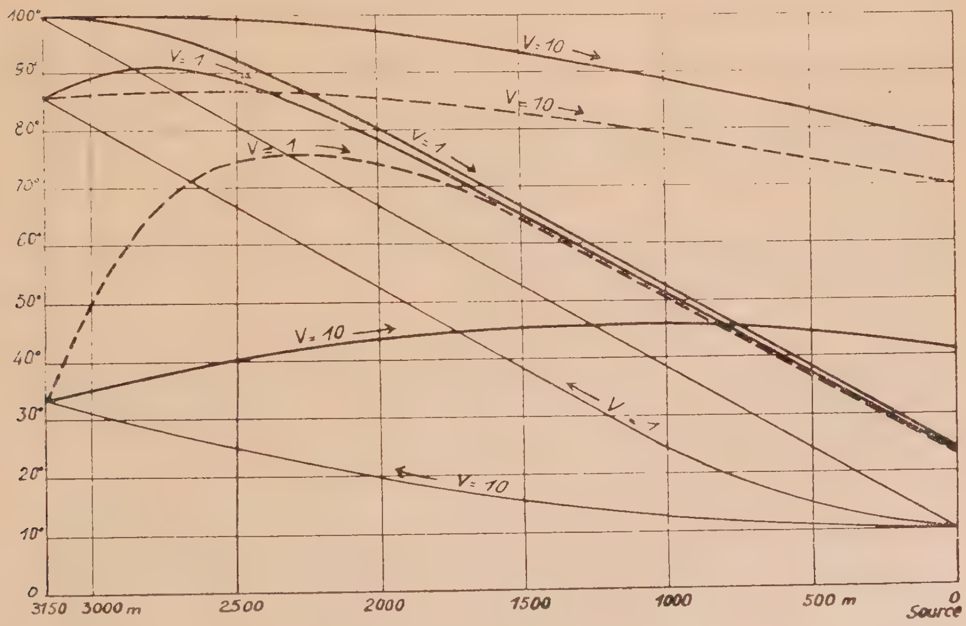

FIG. 3

Ces formules (5), (6) te (7) permettent donc de calculer la température de l'eau ascendante, dans les différentes parties de son chenal de forme cylindrique.

Prenons plusieurs exemples d'un chenal vertical, pour lequel : $G = 1/35000$, $H = 315000$ cm, $\theta^s = 10^\circ$, $K = 0{,}005$ unités C.G.S.. On donne à l'origine $\theta_H = 100^\circ$. Et l'on a posé $\rho = 10$ cm et $\log e\, R_1/\rho = 5$, d'où $R_1 = 1480$ cm.

Premier exemple : à l'origine, la température de l'eau est égale à celle du sol, $\theta_0 = \theta_H = 100^\circ$ et la vitesse de l'eau est $V = 1$ cm/sec.

2e exemple : mêmes caractéristiques, mais $V = 10$ cm/sec.

3e exemple : à l'origine, la température de l'eau, $\theta_0 = 87^e75$, est inférieure à celle du terrain et $V = 10$ cm/sec.

4e exemple : mêmes conditions, mais $V = 1$ cm/sec.

5e exemple : $\theta_0 = 33^e15$ et $V = 10$ cm/sec;

6e exemple : idem, mais $V = 1$ cm/sec.

Cas des chenaux subverticaux ou verticaux descendants.

Ce cas ne diffère du précédent que par une circulation de l'eau de la surface du sol vers la profondeur, l'origine étant située à la surface du sol. Comme précédemment θ_s = température de la zone neutre du sol et θ_0 = température de l'eau à l'origine, c'est-à-dire ici à la surface du sol.

Il nous suffit donc de reprendre les équations (4) et (2) dans lesquelles par suite du changement d'origine, on affectera *h* et V du signe négatif et on remplacera θ_H par θ_s. Il vient alors :

$$\frac{d\theta}{dh} = -\frac{N\theta}{\sin\alpha} + \frac{NGh}{\sin\alpha} + \frac{N\theta_s}{\sin\alpha} \tag{8}$$

En posant : $a = N/\sin\alpha$ $\quad b = NG/\sin\alpha$ et $\quad c = N\theta_s/\sin\alpha$

$$\frac{d\theta}{dh} = -a\theta + bh + c$$

$$\theta = \frac{abh + ca - b}{a^2} + \mu e^{-ah}$$

et avec conditions aux limites $\theta = \theta_0$ pour $h = 0$:

$$\theta = \theta_0 e^{-ah} + \frac{(ca - b)}{a^2}(1 - e^{-ah}) + \frac{bh}{a}$$

d'où

$$\theta = \theta_s + Gh + (\theta_0 - \theta_s)\, e^{-\frac{Nh}{\sin\alpha}} - \left(\frac{G}{N}\sin\alpha\right)\left(1 - e^{-\frac{Nh}{\sin\alpha}}\right) \tag{9}$$

formule donnant la température de l'eau à la profondeur *h*, sous le sol. $\theta_s + Gh$ est la température θ_1 du sol à la profondeur *h*,

$$N = \frac{2K}{cV\rho^2 \log e\, R_1/\rho} \tag{10}$$

Dans le cas de beaucoup le plus courant, la température de l'eau, à l'entrée du circuit, peut être considérée comme égale à la température θ_s de la zone neutre, et l'équation se simplifie et devient :

$$\theta = \theta_s + Gh - \left(\frac{G}{N}\sin\alpha\right)\left(1 - e^{-\frac{Nh}{\sin\alpha}}\right) \tag{11}$$

Dans la fig. 3 nous avons donné deux exemples d'une eau descendant dans un chenal vertical. La température d'entrée de l'eau $\theta_0 = \theta_s = 10^e$. Dans le premier cas, $V = 10$ cm/sec. Dans le deuxième cas, $V = 1$ cm. Comme précédemment, dans les deux cas, $G = 1/35000$, $K = 0{,}005$ unités C.G.S., $\rho = 10$ cm et $\log e\, R_1/\rho = 5$, d'où $R_1 = 1480$ cm.

Cas d'une eau ascendante, circulant dans une couche verticale ou subverticale.

Soit une couche verticale, de longueur infinie dans le sens horizontal (fig. 1). Son inclinaison est α. Un flux vertical de chaleur terrestre, de gradient G, monte de bas en haut, la conductibilité calorifique du terrain étant K. Considérons une portion de cette couche, ayant l'unité de longueur dans le sens horizontal, soit 1 cm. L'origine du système est à la profondeur H sous le sol. Considérons une longueur *dl* infiniment petite de cette nappe, au point d'ordonnée h. La température de l'eau y est θ. A la distance R_1, perpendiculaire à la couche, distance à partir de laquelle les isothermes ne se trouvent plus modifiées par suite des échanges entre l'eau et le terrain, distance par conséquent à partir de laquelle, la température du sol est la température θ_1 donnée par degré géothermique. En réalité il faudrait considérer les températures θ_2 et θ_3 situées aux distances R_1 de part et d'autre de la couche. Nous supposerons qu'il est licite de prendre la température θ_1, du sol à l'ordonnée h, intermédiaire entre θ_2 et θ_3, et d'autant plus licite que la couche est voisine de la position verticale.

Il pénétrera donc le long de dl, par l'ensemble des deux faces, le flux de chaleur suivant, pendant l'unité de temps :

$$P_o = 2\,K\,\frac{\theta_1 - \theta}{R_1}\,dl$$

Soit q le débit de l'eau, c la chaleur spécifique de l'eau et $d\theta$ l'augmentation de température pendant l'unité de temps, on aura donc

$$cqd = 2K\,dl\,\frac{\theta_1 - \theta}{R_1}$$

Et comme $dl = dh/\sin\alpha$, on a alors :

$$\frac{d\theta}{dh} = \frac{2\,K}{c\,q\,R_1\sin\alpha}(\theta_1 - \theta)$$

et en posant

(12) $$N = \frac{2\,K}{c\,q\,R_1}$$

on a :

$$\frac{dh}{d\theta} = \frac{N}{\sin\alpha}(\theta_1 - \theta)$$

d'où pour $h \neq 0$:

(13) $$\theta = \theta_H - G\,h + (\theta_o - \theta_H)\,e^{-\frac{Nh}{\sin\alpha}} + \left(\frac{G}{N}\sin\alpha\right)\left(1 - e^{-\frac{Nh}{\sin\alpha}}\right)$$

dans laquelle, θ, θ_H, $\theta_H - G\,h$, G, α, K et c ont même signification que dans la formule 5,

(14) $$N = \frac{2\,K}{c\,q\,R_1}$$

q est le débit à travers une section ayant une unité de longueur de la couche dans le sens horizontal, $R_1 =$ la distance à partir de laquelle, les isothermes du flux de chaleur ne se trouvent plus modifiés.

Lorsque la température de l'eau θ_o à l'origine, à la profondeur H, est égale à celle θ_H du terrain, on a donc :

(15) $$\theta = \theta_H - G\,h + \left(\frac{G}{N}\sin\alpha\right)\left(1 - e^{-\frac{Nh}{\sin\alpha}}\right)$$

Cas d'une eau descendante, circulant dans une couche verticale ou subverticale.

Il est aisé de voir que l'on obtient ici une formule identique à celle d'un chenal descendant :

(16) $$\theta_o = \theta_s + G\,h + (\theta - \theta_s)\,e^{-\frac{Nh}{\sin\alpha}} - \left(\frac{G}{N}\sin\alpha\right)\left(1 - e^{-\frac{Nh}{\sin\alpha}}\right)$$

dans laquelle, les lettres ont même signification, sauf N, car $N - \frac{2\,K}{c\,q\,R_1}$ comme précédemment, équations (12) et (14).

Lorsque la température θ_o de l'eau, à l'entrée, est égale à celle θ_s de la couche neutre du sol :

(17) $$\theta = \theta_s + G\,h - \left(\frac{G}{N}\sin\alpha\right)\left(1 - e^{-\frac{Nh}{\sin\alpha}}\right)$$

Remarques générales sur la thermique des eaux descendant et remontant dans des chenaux cylindiques ou des couches, les uns et les autres verticaux et subverticaux.

Prenons le cas d'un circuit descendant, puis ascendant. Examinons d'abord la thermique des chenaux ou des couches à eaux descendantes.

D'une manière générale, à la surface du sol, l'eau entre à une température égale ou approximativement égale à celle de la zone neutre du sol. Cette dernière est d'ailleurs très voisine de la température moyenne annuelle de l'air du lieu considéréré. Elle lui est de 0°8 dans les climats humides à 1°8 dans les climats très secs supérieure.

C'est dans ces branches descendantes des circuits souterrains que se fait surtout l'absorption des calories. Autrement dit l'eau y a une action nettement refroidissante sur le terrain. A partir du sol, la température de l'eau s'accroit de plus en plus rapidement. Mais dès une certaine profondeur, l'accroissement tend à rester le même et égal à l'accroissement de la température du terrain en fonction de la profondeur; de telle sorte qu'il reste, au bout de ce temps, un écart presque constant de température de l'eau et celle du terrain. Autrement dit, à partir d'une certaine profondeur, l'accroissement de chaleur de l'eau reste constant.

Mais dans cette branche descendante, l'eau se réchauffe d'autant moins vite, et la profondeur à partir de laquelle l'accroissement de température de l'eau suit à peu près l'accroissement de température du terrain est d'autant plus grande, que la vitesse de circulation de l'eau, c'est-à-dire son débit, est plus grande. L'eau arrive donc froide à la fin de sa descente, et d'autant plus froide que le débit est plus grand.

A l'origine de la branche ascendante, dans un tel circuit imaginé, l'eau est donc plus ou moins froide que le terrain. Dans une première partie de ce circuit ascendant, l'eau continue à recevoir des calories. Et elle se réchauffe jusqu'à ce que la température de l'eau atteigne celle du terrain. En effet au point où l'eau atteint son maximum de température $\frac{d\theta}{dh} = \mathrm{N}\,\frac{\theta_1 - \theta}{\sin\alpha} = 0$. On a donc bien $\theta = \theta_1$. La température de l'eau est la même que celle du terrain encaissant.

L'ordonnée *h* au-dessus de l'origine, et à laquelle cette température maxima a lieu, est donnée par les équations (5) ou (13), suivant le cas, et dans lesquelles on aura rendu égal $\theta = \theta_\mathrm{H} - \mathrm{G}\,h$, puisque là $\theta_\mathrm{H} - \mathrm{G}\,h = 0$. Mais la profondeur H sous le sol à laquelle cette température maxima a lieu, peut être calculée à l'aide de la formule :

$$\theta = \theta_\mathrm{S} + \left(\frac{\mathrm{G}}{\mathrm{N}}\sin\alpha\right)\left(1 - e^{-\frac{\mathrm{N}h}{\sin\alpha}}\right)$$

dans laquelle θ est la température de la source.

A partir de ce point, l'eau se refroidit, d'abord lentement, puis plus rapidement jusqu'à une certaine profondeur au-dessus de laquelle la diminution de température a tendance à être la même que celle du terrain. Mais l'eau reste constamment plus chaude que le terrain. De telle sorte qu'elle arrive à la source avec une température bien supérieure à celle de la zone neutre du sol. Et la source sera d'autant plus chaude que la vitesse ou le débit de l'eau seront plus grands et que la température de l'eau à l'origine est plus élevée.

Pour des débits ou des vitesses extrêmement faibles, la température de l'eau a tendance à se rapprocher de celle du terrain et d'autant plus et d'autant plus rapidement que ceux-ci sont faibles; de telle sorte que la température de l'eau d'une source de petit débit, peut devenir très voisine de celle de la zone neutre du sol.

Il est intéressant de remarquer, et ceci peut être très important en ce qui concerne les dissolutions et les réactions chimiques à l'intérieur du circuit, que ce ne sont pas les eaux des sources les plus chaudes, à gros débit, qui dans leur trajet souterrain ont, à un moment donné, la température la plus élevée, mais les sources à faible débit et relativement froides; car la profondeur de la température maxima est d'autant plus grande que la vitesse ou le débit de l'eau est plus faible, comme on peut le voir d'après nos formules ou les graphiques.

Dans un tel circuit, on le voit, l'eau se réchauffe, non seulement dans sa partie descendante, mais aussi dans la première partie de sa branche ascendante, première partie d'autant plus courte que le débit est plus petit. Et ce n'est que dans la deuxième partie de la branche ascendante que l'eau se refroidit.

Dans le cas où l'eau est à l'origine de la branche ascendante à la même température que le terrain, ce qui peut se produire après une longue circulation horizontale ou subhorizontale, l'allure de la courbe de température est la même que la partie qui suit le maximum des cas précédents.

Le flux de chaleur terrestre et la température des eaux thermales.

Lorsque l'on étudie la thermalité d'un système aquifère, il ne s'agit pas seulement de rechercher si l'eau peut provenir d'une zone assez chaude, c'est-à-dire provenir d'une zone à potentiel thermique assez élevé pour lui donner sa température. Mais il s'agit également de savoir si l'eau peut recevoir un flux de chaleur suffisant, pour maintenir constante son alimentation en calories. C'est ce que l'on oublie de faire. Or ce flux est le flux terrestre. Et nous le connaissons, il est de 1,8 à $1{,}6 \times 10^{-6}$ calories gr /cm^2 sec.

La section de flux de chaleur terrestre, captée par une source thermale à faible débit est petite. Ainsi le nombre de calories gr remontées par seconde par les sources suivantes (1) est : Vichy Grande Grille 20900 ; Vichy Puits Carré 60400, Chaudesaigues 351000, Eaux Bonnes (B. Pyr.) 14000. La section de flux de chaleur terrestre, détournée par ces différentes sources est ainsi respectivement de 1,16 — 3,33 — 6,16 — 0,78 km^2. L'échauffement de l'eau, lors de la descente et d'une partie de la remontée de celle-ci dans des veines relativement étroites, de faible section, peut ainsi facilement s'expliquer par un tel détournement du flux de chaleur entourant les veines aquifères.

Mais lorsqu'il s'agit d'une source thermale à très gros débit, et surtout si cette source est à température très élevée, il n'en va pas de même. Ainsi les sources de Dax (France), à température de 64° à la Nèhe, débitant 10000 m^3 /jour, sur un front de 1600 m, hors d'un banc presque vertical de calcaire dolomitique du Sénonien de 100 m de largeur, remontent 71410.000 cal gr /sec. En admettant une température de la zone neutre du sol égale à 14° et en supposant donc que ces eaux ont une origine météorique, elles empruntent alors 5.790.000 cal gr /sec au flux terrestre. La surface de ce flux détourné doit alors être de 320 à 360 km^2.

Un exemple : les sources de Dax.

Nous allons prendre comme exemple les sources de Dax dont il vient d'être question.

Or les auteurs (2) et (3), qui ont étudié l'origine de cette eau, ont pensé qu'elle provenait de l'infiltration de l'Adour et même (3) particulièrement de l'infiltration dans le banc de calcaire dolomitique sénonien de Dax même. Après être descendue presque verticalement jusqu'à 1800 à 2000 mètres de profondeur, l'eau remonterait chaude, également presque verticalement dans ce même banc calcaire. Il se réaliserait ainsi un thermo-siphon, ce qui a priori est possible. L'argument invoqué par ces auteurs est que le débit des sources augmente lorsque le niveau de l'Adour s'élève (4). Or l'eau est supposée monter et descendre verticalement dans ce banc de calcaire dolomitique, dans une section de $1{,}6 \times 0{,}1 = 0$ km^2,16. On voit donc les disproportions énormes qui existent entre cette section et les 320 à 360 km^2 représentant la section de flux de chaleur terrestre nécessaire pour échauffer l'eau. Une section de 10 km de rayon de flux, autour de la source, serait ainsi entièrement détournée. Et la modification de forme du flux autour de ce faisceau capté doit nécessairement s'étendre beaucoup plus loin autour des veines aquifères. Cela parait bien improbable pour une eau ne remontant que de 1800 à 2000 mètres de profondeur. Il vaudrait donc mieux imaginer une chauffe de l'eau, dans une région synclinale par exemple.

Examinons maintenant la thermique des branches ascendante et descendante.

La branche ascendante fournit des résultats plus ou moins satisfaisants dans de très larges conditions géophysiques, car c'est dans cette branche ascendante ou du moins dans la partie la plus élevée de celle-ci, que se fait la dissipation de la chaleur. Mais ces résultats ne sont pas caractéristiques. Admettons, par exemple que l'eau monte sur une longueur de 1 km ($q = 1$ cc /sec, 157 par cm), que $G = 1/35000$. Donnons à R_1 différentes valeurs, la profondeur h_m à laquelle l'eau est à la même température que celle du terrain et le refroidissement $\Delta\theta$ de l'eau à partir de ce point, jusqu'à la source, sont donnés dans le tableau suivant :

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(1) L. de Launay Recherches, captage et aménagement des sources thermominérales, Paris, 1899, p. 168 et suiv. Ces chiffres seraient encore plus petits si, admettant l'origine météorologique de ces sources,on déduisait alors la température de la zone neutre du sol de la température de la source.

(2) H. Delfour. — Contribution à l'étude physicochimique des eaux thermales de Dax et de leurs variations en fonction des crues de l'Adour, (thèse, Fac. Médecine de Bordeaux, 1 br., 94 p., Bordeaux, 1937.

(3) P. Rhodain. — Les venues hydrominérales et la genèse des gisements minéraux en Aquitaine., *Ann. des Mines* (partie tech.), (14), Mém. 1, p. 5-10, 1942.

J. Dupouy-Camet. — Remarques hydrogéologiques sur les eaux thermominérales de Dax et de la région dacquoise., *Ann. Hydrologie et Climatologie*, XVII, n° 62, p. 116-130, 1946.

(4) Remarquons que l'influence des crues de l'Adour sur le débit des sources pourrait tout aussi bien s'expliquer par l'exercice d'une contre-pression sur les griffons, contre-pression diminuant les pertes d'eau thermale dans les alluvions, les forçant à sortir aux griffons (cf. le principe de François).

R_1	16000 m	1000 m	100 m	50 m	30 m	20 m
h_m	1760 m	1770 m	1900 m	2070 m	2430 m	3320 m
$\Delta\theta$	0°3	0°6	4°3	9°1	19°4	44°9

Il est très vraisemblable que R_1 mesure plus de 30 mètres. On voit alors que si cela est bien ainsi, les erreurs que l'on peut commettre sur les appréciations doivent être petites.

Mais il n'en est pas de même lorsque l'on étudie la thermique de la branche descendante, car pour R_1 il faut prendre une valeur au moins aussi grande, sinon plus grande, que le rayon du flux intercepté. Prenons par exemple, le cas de R_1 très grands. Il vaut mieux alors assimiler la descente de l'eau à celle qui aurait lieu dans une conduite cylindique dont la section aurait une surface égale à celle de la zone d'infiltration à travers le banc calcaire, soit 40700 m^2 (1000 × 40m,7 pour la commodité du calcul). On a alors $\rho = 10900$ cm. Donnons à R_1 les valeurs de 10 et 1 km (exprimées en cm.). On a alors les températures suivantes de l'eau aux diverses profondeurs ci-dessous, avec $\theta_s = 14°$:

	R_1			R	
h	10 km	1 km	*h*	100 m	10 m
1000 m	14°15	14°18	1157	15°6	26°2
2000	14°36	14°70	2314	20°2	51°6
			2800		64°
3500	15°04	15°86	3000		69°7

En réalité R_1 devrait être de bien plus de 10 km, étant donné le flux de chaleur nécessaire à l'alimentation de l'eau en calories. Or déjà avec $R_1 = 10$ km, nous constatons qu'à 3500 m l'eau ne s'est réchauffée que de 1°04. Le système hydraulique ainsi envisagé ne peut donc expliquer le réchauffement de la branche descendante. En effet même avec $R_1 = 100$ m, une remontée à partir de 2314 m où l'eau a une température de 20°2, l'eau de la source n'aurait qu'une température de 26°1, au lieu des 64°.

Pour qu'il y ait réchauffement, la densité de flux de chaleur doit donc être beaucoup plus grande, par conséquent R_1 beaucoup plus petit. Avec des R_1 inférieurs à 1 km, il est nécessaire de prendre la valeur de N correspondant au mouvement dans un banc vertical de grande longueur. Soit la descente sur une longueur de 1 km. On a alors avec $R_1 = 100$ m et $R_1 = 10$ m (exprimés en cm) les températures indiquées au tableau précédent. On constate donc qu'il faudrait un flux de chaleur très dense tel que R_1 soit voisin de 10 m. On ne peut alors admettre un réchauffement de l'eau de l'Adour dans la branche verticale descendante du siphon.

Lorsque l'on étudie l'origine des eaux chaudes par la thermalité, il est donc nécessaire de tenir compte non seulement du potentiel thermique, mais aussi du flux de chaleur capté. Et il importe également d'examiner non seulement la thermique de la branche ascendante, mais surtout celle de la branche descendante.

2) LES SOURCES THERMO-MINERALES DE L'ALGERIE

par

S. GUIGUE et G. BETIER

D'après les documents publiés à diverses époques par le Service des Mines, il existe sur le territoire de l'Algérie 175 sources thermo-minérales d'importance très variable. Leur nombre croît régulièrement quand on se déplace de l'Ouest vers l'Est.

Le nombre des gisements métallifères reconnus varie dans le même sens.

Situation géographique	Nombre de concessions minières	Nombre de sources thermales	% approximatif
Province d'Oran	15	17	10%
Province d'Alger	27	35	20%
Province de Constantine	98	122	70%
	140	174	

Gisements miniers et sources thermales sont plus fréquents dans l'Atlas Tellien, zone de plissements très intenses, que dans l'Atlas Saharien, de structure beaucoup plus simple et plus régulière.

Le tableau ci-joint récapitule, d'après les travaux réalisés par S. Guigue de 1937 à 1947 (1) les résultats d'analyses et les températures à l'émergence des 35 groupes de griffons les plus importants, rangés d'après leurs propriétés chimiques fondamentales, qui déterminent leur « physionomie ».

La concentration des sels dissous est sans rapport apparent avec la température mesurée à l'émergence. Celle-ci varie de 19° (Ben-Haroun) à 95° (Hammam-Meskhoutine). Le poids de l'extrait sec par litre, mesuré à 180°, varie de 0 gr 37 (Hammam-Bradaa 28°) à 59 grs 5 (Aïn-Mentila 35°). Les eaux qui surgissent à la température la plus élevée (Hammam-Meskhoutine) ne renferment qu'un gramme et demi de sels dissous par litre.

On vérifie ainsi le fait bien connu que la minéralisation des eaux est déterminée surtout par la nature chimique et minéralogique des sédiments qu'elles traversent et, accessoirement, par leur vitesse de circulation souterraine.

Les eaux les plus minéralisées (Aïn-Mentila 59 grs 5 par litre — Hammam-Melouane 29 grs 42 — H. Beni-Guecha 16 grs 87 — H. el Biban 15 grs 4 — H. Salahine 9 grs — H. Ksenna 5 grs 5) sont en relation directe certaine ou très probable avec les sédiments gypso-salins du Trias, si répandus en Algérie où ils ont donné lieu à des phénomènes de diapyrisme particulièrement accentués.

On peut extrapoler cette hypothèse sans grand risque à la plupart des eaux sulfurées, chlorurées et sulfatées calciques étudiées. L'échelle des températures observées permet alors dans une certaine mesure, d'apprécier l'intensité des efforts orogéniques dans la zone considérée.

Si l'on rapproche de cette première étude chimique comparative d'ensemble,

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(1) S. Guigue : Les sources thermo-minérales de l'Algérie. Etude géochimique. *(Bulletin du Service de la Carte Géologique de l'Algérie,* 3^{me} série, 2 volumes publiés respectivement en 1940 et 1947).

				Grammes par litre	MILLI		
			Tempé-rature	Extrait à 180°	Na	K	Ca
FACIÈS ALCALIN	S. Leblanc	bicarbo-natées sodiques	24°	1,610	292,8	110,0	183,2
	Takitount		21°7	2,210	614,7	6,0	212,8
	Ben-Haroun	sulfatée sodique	19°	3,312	693,2	11,8	463,4
FACIÈS TERREUX	H. Bou-Hanifia ..	bicarbo-natées calciques	65°	1,314	251,1	30,0	170,0
	Aïn N'Sour ...		18°	0,288	10,3	4,2	59,2
	Sidi-M'Cid ...		29°	0,778	90,4	9,3	138,0
	Le Hamma ...		34°	0,729	74,2	2,9	131,2
	Dj.Lekhal (Aïn Tinn)		31°5	0,553	25,6	5,05	110,7
	O. Hammimime .	sulfatées calciques	46°5	2,391	103,2	13,6	502,4
	H. Guergour		44°3	3,521	347,5	10,3	601,0
	H. Righa		54°	2,466	214,2	15,0	477,2
	O. Ghalia (Béni Hindel) ...		40°	2,444	641,3	14,7	194,8
	H. Melouane		39°5	29,422	10.260,4	120,0	832,5
	H. Bou-Hadjar .	chlorurées	31 à 70°	3,516	845,0	56,6	330,0
	A. Ouarka		46°5	5,609	1.330,6	64,9	466,0
	H. Bradaa		28°	0,371	15.4	traces	89,6
	Les Abdellys ...		34°	0,405	21,9	4,4	82,1
	H. N'Baïls		42°	5,839	1.526,5	106.7	500,0
EAUX SULFURÉES	H. Salahine	sulfurées sodiques	43° —	9,159	2.958,0	90,48	357,0
	H. Berrouaghia ..		38°	1,361	432.3	11,1	64,9
	H. Boughara ...		45°	0,405	52.8	5,2	59,2
	H. Meskoutine ..	sulfurées calciques	80 à 96°	1,466	204.9	46,8	202,4
	Youks-les-Bains .		34°	0,431	20.9	2,51	92,4
	H. Amamrhas ..		58 à 65°	2,190	630,9	22,4	132,4
	H. Bou-Akkaz ...		39°5	2,724	738,3	13,3	236,1
	H. Grous		37°6	1,160	131,9	7,6	191,8
	H. Ksenna		50 à 52°	5,466	1.773,8	37.0	160,1
	H. Bou-Sellam ..		40 à 45°	1,399	208.3	10,6	187,3
	Nazereg (Ouled Khaled)		49°	1,833	314.0	18,5	240,2
	A. Mentila		33°	59,522	21.502,5	143,8	969,8
	H. Zaïd	chloro-sulfurées	39°5	1,017	167,4	15,9	163,2
	H. Tassa		40°6	1,992	410,3	26,2	284,0
	H. El-Biban		80°	15,435	5.122,2	60,8	572,0
	H. Ben Chiguer ..		32°5	3,075	922,9	10.2	153,7
	H. Beni Guécha .		55°	16,876	5.362,2	83,3	934,6

GRAMMES PAR LITRE

Mg	Cl	SO^4	CO^3	SiO^3	Na^2S	H^2S libre	As
57,9	237,8	52,5	653,2	12,3			
36,8	152,6	55,0	1.084,2	21,2			
56,2	550,2	618,0	887,5	31,6			
36.8	372,7	81,3	330,0	47,5			
24,1	19,5	30,2	132,1	12,1			
40,1	145.5	93,2	240,1	20,2			
42,0	120,7	104,8	233,1	20,2			
32,0	39,0	158,4	159,0	23,6			
85,9	78,1	1.367,2	200,0	39,9			
92,5	482,9	1.682,1	167,4	46,3			
31,2	337,2	1.074,9	161,3	55,8			
35,8	951,4	430,7	154,3	24,0			
79,2	15.850,7	1.921,9	325,0	31,6			
49.7	1.608,1	53,9	463,0	75,8			
100,8	2.215,2	1.292,6	68,4	51,8			
23,5	27,6	47,9	154,1	12,6			
32,6	42,6	12,9	191,6	16,7			
77,8	2.733,5	444,9	429,3	20,2			
61,3	3.244,7	1.424,0	962,9	50,5	4,24		
8,8	319,5	197,2	281,7	45,8	3,2		
31,8	56,8	18,9	167,0	13,2	3,12		
37,4	326,6	381,5	183,0	83,9		6,80	0,45
33,5	28,4	45,9	191,7	16,1		1,36	
23,6	880,4	215,4	224,7	60,6		1,70	
41,1	1.153,7	228,8	278,4	34,1		0,61	
34,8	209,4	294,1	203,0	43,4		0,58	
69,4	2.584,4	664,9	155,7	20,7		58,00	
42,8	284,0	456,1	120,2	51,9		8,50	
45,6	415,3	598,2	150,0	50,9		5,78	
232,7	33.015,0	3.600,8	30,6	26,5		159,80	
35,6	333,7	47,7	228,3	23,2		13,60	
36,3	724,2	25,1	451,2	25,2		6,80	
95,7	8.278,6	1.002,1	200,0	85,1		14,90	
52,5	1.476,8	193,6	225,1	40,4		0,17	
103,3	8.555,5	1.163,5	580,3	93,5		1,56	

les analyses pratiquement utilisables (1) de 27 autres sources énumérées dans l'ouvrage du Dr Hanriot (2) on peut classer approximativement comme suit les soixante sources les plus importantes de l'Algérie :

— Alcalines	7
— Bicarbonatées calciques	10
— Sulfatées calciques	12
— Chlorurées	6
— Sulfurées	27

On notera tout particulièrement la proportion élevée des sources sulfurées.

A l'exception de la présence d'arsenic dans les eaux d'Hammam-Meskhoutine (95°) aucun argument d'ordre chimique ne semble pourvoir être invoqué en faveur d'une relation directe entre le volcanisme et les sources thermales de l'Algérie.

Le Laboratoire du Service des Mines de l'Algérie poursuit méthodiquement l'étude des propriétés chimiques et physicochimiques de ces eaux qui jusqu'ici n'ont pas retenu suffisamment l'attention des chercheurs.

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(1) C'est-à-dire les analyses pour lesquelles les erreurs apparentes sont faibles.
(2) M. Hanriot, *Les eaux minérales de l'Algérie*, chez Dunod et Pinat — 1911, Paris.

3) ON THE MINERAL HOT SPRINGS OF THE EUGANEAN HILLS AND PARTICULARLY OF ABANO (PADOVA)

by

DAL PIAZ G.

The Euganei form a modest group of hills that rises almost in the centre of the Venetian plain. They correspond to an eruptive group which acted in the Tertiary Era originating « trachitiche » domeshaped intrusions, « liparitiche » and base castings and relative projections of « piroplastici » materials which gave birth to tuffs.

A stripe of sedimentary soils, however broken and discontinuous, of the Eocene, Cretaceo and some localities of the Giurasse, which constitue the remains of the ancient sedimentary cover which is broken and raised specially in the domeshaped intrusions, is found around and astride the small volcanic center.

It is easy to guess that the presence of thermomineral waters, for which the Euganean hills are well-known, is directly connected to the existence of this ancient volcanic group whose fundamental reservoir, however far from the prior endogenous activity, acts still as a centre of thermal influence.

In fact it is known that every active or extinguished volcanic centre, in relatively late geologic ages, gives out, mainly when there are lavic castings of the base zone, emanations of gases, vapours and water at high temperature. Such phenomena are usually called activity of « fumarola ».

The composition and temperature of such gases and vapours are very different according to the age and constitution of the eruptive center from which they derive.

When the temperature of the emission materials of the « fumarole » is very high (even 500 degrees) the aqueous vapour cannot condense exceeding its critical point and we have dry « fumarole ». When, on the contrary, the temperature of the « fumarole » gaseous materials is not very high, only about 100 degrees or even less, the aqueous vapours condens giving origin directly to springs. May be that a part of the vapours of warmer (dry) « fumarole », because of their cooling during their underground way or because their getting in contact with hydraulic cool layers, condenses, while another part keeps at the aeriform state accompanying the gushing out of the springs.

The aeriform manifestations which accompany the Euganean thermal springs are composed mainly of carbonic acid, nitrogen, helium, sulphurous anhydride sulphydric acid and quantities more or less abundant of aqueous vapour. The sodium-chlorides and the calcium mixtures are specially abundant among the salts carried by these springs and this is in relations to the fundamental constitution of the eruptive rocks of the Euganean group.

A matter that was long discussed by geologists and by geochemists, among wich we mention **Sainte Claire Deville**, **Fouqué**, **Suess**, **Daubré**, **Wolf**, **Gautier**, **Hellen**, **Haug**, **Ponte**, **De marchi** etc. is if thermomineral waters in general owe their origin to the condensation of the vapours which accompany the emissions of the « fumarole » such as those exuded from the rocks of the ancient volcanic reservoir, (*aquae juveniles*) or if they are only waters of superficial penetration, or better waters of the common deep hydric layers (*aquae vadosae*), which are thermalized and mineralized by getting in contact with the different manifestations of post-volcanic activity such as : gases, vapours and waters at high temperature coming out from the flaws of the eruptive rocks of the castings and of the intrusions.

In the particular case of the Euganean hills, the interpretation which attributes exclusively thermomineral waters origin to a sort of distillation of deep eruptive rocks of volcanic reservoir, however qualitatively supported by laboratory tests, cannot justify the great quantity of water of the various thermal springs of the region, a quantity that reaches the ten millions of cubic meters a year.

Though we admit that a certain contribution to the increase of the Euganean thermal springs may derive either from the direct condensation of the vapours emanated by the eruptive rocks or from the penetration of meteoric waters warmed in depth and then pushed to the surface by the tension of gases and vapours, the existence and the abundance of Euganean thermal springs may be explained substantially by other facts connected to causes that for their entity can give better reasons of the great phenomenon.

The true thermo-mineral springs gushing out spontaneously from the soil are relatively few. Among them we mention those of Montirone, of Montegrotto, of Battaglia, of S. Elena, of the Bagnarola, of S. Bartolomeo and some others of less importance.

All the remaining ones, and there are very many ot them, were originated drilling at different depths the sedimentary cover. This happens on a large area and shows, by a long never failed experience, that the various drillings made at Abano and in other Euganean localities, are alimented by a sort of common source formed of a large zone of hydric thermomineral impregnation, which exists at a certain depth of the whole region.

How this useful natural water deposit has been formed and maintained, and how this water is thermalized and mineralized, may be explained only going back to the fundamental laws of underground hydric circulation in alluvial soils and to the local existence of an extinguished volcanic centre in a relatively recent geological age. In fact, we know that the waters of the common hydric layers which are originated by the alluvial imbibition of the Vicentina and Trevigiana high plain, flow slowly towards the South. During their slow underground way they succeed to touch at least in part, the fundamental foot of Euganean hills, then get in contact with those factors of endogenous activity (gases, vapours, warm waters) corresponding to the manifestations of a volcanic reservoir extinguished in its superficial function, but still in a certain activity of secondary importance in its deep parts. In this way the thermo-mineral waters of all the region surrounding the Euganean hills are originated. Such layers (as Wolff and Hellen demonstrated for other countries) must be considered as common hydric layers which were afterwards thermalized by the mentioned factors.

These conclusions are strengthen by the fact that the deep explorations till now fulfilled in the Euganean region show how starting from Montirone, which is the centre of thermal manifestations of Abano, and proceeding towards the North (from where normal hydric layers descend), the temperature of thermo-mineral waters decreases quickly while it increases proceeding towards the South and then it gets lower again beyond a certain limit, where probably endogenous actions are no more felt.

One would know why, on the contrary of what happens in many other modern and ancient eruptive centres, in which the « fumarola » substances are emanated directly from the surface splits of the eruptive rocks along the sides or at the foot of the volcanic group, in the Euganean hills the emission mouths of these thermalizing substances must be buried under powerful alluvial covers.

The fact is easily explained if we think that during a great part of the quaternary Era, the Venetian plain did not exist and the Euganean hills raised through a bottom of a marine and afterwards of a lagoon type, in continuation to the Adriatic sea. The alluvions caused by rivers and torrents little by little filled up this valley so that the base of Euganean hills and with it the mouths of emanation of thermalizing factors remained buried. However it is enough to drill the alluvial cover to provoke immediately the going out of warm water as the numerous terebrated wells in the area around Montirone springs, testify.

One might think that the thermal waters of the Euganean springs are originated instead from the superficial absorption by the small hilly groups of rain falls which, penetrating in depth, would be afterwards thermalized. However, also in this case, they would be vadose waters, warmed by the residue heat of the ancient volcanic reservoir. Anyhow, considering the small extension of the Euganean Hills, one could not be convinced how a modest seasonal absorption, inevitably exiguous, because effected through impermeable strata, can furnish such great quantity of waters as that one gushing out from the various thermal springs and by the very numerous bored wells.

Besides if this is really the origin of the thermal waters of Abano and the regions surrounding it, how to explain the constancy of their discharge considering the seasonal variations of rain falls from which the supplies and the action of the springs depend ? The late periods of exceptional dryness would furnish very significative data about it, while the repercussion on the discharges was almost trifling and proportionated to a layer action. The water of penetration by inflitration through the massive of the Euganean Hills, thermalized afterwards in depth by the residue heat of the volcanic group, can cause a certain increase to the thermo-mineral Euganean sources, but it cannot by itself give a reason of the entity of the hydrologic phenomenon which is of wider proportions.

At last we think that like the very numerous Artesian wells of the Venetian plain which draw their potable water from the deep hydric layers descending along alluvial cones of dejection, the tubular wells of Abano draw water from a similar hydric layer, which however getting in contact with the base of the eruptive Euganean centre in the stage of late activity (consisting in emanation of gases, vapours and water at high temperature) raise their temperature and are enriched by various mineral substances.

Natural springs, that are not provoked by drillings, such as those of Montirone, Monteortone, etc. are but some veins deriving from the same thermalized layer, whose waters, because of the tension of the gases and vapours which accompany them, have succeeded to open spontaneously a way through the sandy and clayish sediments constituting the alluvial cover of the plain.

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4) LES MODIFICATIONS DE LA COMPOSITION CHIMIQUE DE L'EAU DANS UNE MEME NAPPE

par

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La composition chimique d'une eau souterraine est quelque chose de très vivant. Et on aurait bien tort de ne la considérer qu'au point de vue statique.

Dans une même nappe, l'eau peut subir les effets de plusieurs phénomènes : une concentration, des échanges de bases, une réduction des sulfates, capables de modifier partiellement et parfois du tout au tout l'aspect chimique de l'eau.

La concentration peut se faire par évaporation ou par dissolution. Dans la concentration par évaporation, il y a une influence essentiellement climatérique. L'évaporation a en effet surtout lieu dans les zones d'alimentation des nappes. L'eau de pluie ayant imbibé le sol, est à nouveau reprise par évaporation, dans la zone d'évaporation du sol. L'eau du sol se concentre ainsi progressivement. La pluie suivante, permettant une infiltration profonde, entraîne, plus ou moins diluées, ces solutions concentrées du sol jusqu'aux nappes d'eau. On conçoit aisément que plus l'intervalle entre deux pluies pénétrantes, alimentaires est grand, que, moins les pluies sont abondantes, que plus la température et le déficit de saturation de l'air sont élevés, plus alors les eaux d'infiltration se trouvent être concentrées. C'est ce qui explique que la concentration des eaux souterraines augmente des régions tempérées jusqu'aux régions tropicales, pour diminuer ensuite vers les régions équatoriales (1).

Il est toutefois à noter qu'il peut y avoir concentration par évaporation dans les nappes profondes, lorsque des gaz s'échappent de ces nappes, les gaz entraînant alors de la vapeur d'eau. C'est le cas en particulier des eaux de certains gisements pétrolifères d'où partent des hydrocarbures gazeux et du gaz carbonique (2).

En ce qui concerne la concentration par dissolution, les facteurs principaux sont la température, la pression, la grandeur de la surface de contact, le volume de l'eau en présence, le temps. Et l'on sait, d'après la loi de Nernst, que « la vitesse de dissolution d'un corps solide est proportionnelle au déficit de saturation ». Par conséquent plus une nappe sera profonde, c'est-à-dire plus son eau aura une température élevée, plus le terrain aquifère sera à grain fin dans le cas de porosité d'interstice, ou sera diaclasé dans le cas de terrain à porosité de fissures, plus les pores ou les fissures seront étroites, plus la vitesse de circulation de l'eau sera faible et plus la nappe sera longue, plus alors les eaux seront concentrées.

La concentration par dissolution ne peut évidemment dépasser un certain degré, car les eaux tendent à être en équilibre physico-chimique avec les roches dans lesquelles elles circulent. Cet équilibre, lui-même, ne tend à être atteint qu'après un temps plus ou moins long — et il ne l'est complètement qu'au temps infini —temps plus ou moins long suivant la nature du terrain et suivant la concentration existante des sels dans l'eau, conformément à la loi de Nernst.

Examinons maintenant quelles sont les principales modifications de la composition chimique, résultant de la concentration par dissolution. Les principaux radicaux contenus dans les eaux sont les suivants : Ca, Mg, Na, Cl, SO_4, CO_3, HCO_3. Nous avons donc à examiner les solubilités et les produits de solubilité des sels résultant de la combinaison de ces divers ions. Ces sels peuvent être rangés suivant leur solubilité, S, en grammes dans un kilog d'eau :

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(1) H. Schoeller. — L'influence du climat sur la composition chimique des eaux souterraines vadoses *Bull. Soc. géologique de France*, (5), XI, 267-289, 1941.

(2) R. van A. Mills et Roger C. Wells. — The evaporation and concentration of waters associated with petroleum and natural gas. U.S. geological Survey, *Bull.* 693, 104 p., 4 pl., 5 fig., 1919.

	S	Produit de solubilité
$CaCO_3$		0,48 × 10^{-8} à 25° C
$CaSO_4$	2,016 à 18° C	6,1 × 10^{-5} à 18°
$MgCO_3, 3H_2O$		1,4 × 10^{-4} à 16°
$NaHCO_3$	96 à 20°	
Na_2SO_4	193 à 20°	
Na_2CO_3	213 à 20°	
NaCl	263 à 20°	
$MgSO_4$	355 à 20°	
$MgCl_2$	546 à 20°	
$CaCl_2$	745 à 20°	

D'un autre côté, les principaux sels que l'on rencontre dans les roches et qui peuvent par conséquent être dissous en grande quantité, sont $CaCO_3$, $CaSO_4$, $MgCO_3$ et NaCl. Les autres ne se trouvent qu'à l'état de traces ou tout à fait exceptionnellement dans certains gisements. Déjà nous pouvons nous rendre compte de ce que certains sels ne pourront jamais être précipités hors des eaux souterraines vadoses, par exemple $NaHCO_3$ et Na_2CO_3. Il faudrait en effet avoir dans ces eaux souterraines d'origine météorique des valeurs de CO_3 combiné, de 2500 à 7500 mg par litre au moins, valeurs qui n'y sont jamais atteintes.

Les eaux souterraines, sauf celles de terrains exclusivement siliceux ou silicatés sont déjà sensiblement saturées en carbonate et en bicarbonate de chaux. D'un autre côté, la tension du gaz carbonique des eaux souterraines reste toujours à peu près le même (1), celle du gaz carbonique de l'atmosphère du sol agricole, c'est-à-dire comprise entre 0,005 et 0,06. Il ne peut donc y avoir, par suite de la concentration par dissolution, une augmentation de la teneur en carbonate et en bicarbonate, sauf évidemment si les eaux se chargent en ions autres que Ca, CO_3 et HCO_3, ce qui a pour effet d'augmenter le produit de solubilité, mais jamais dans de grandes proportions. D'un autre côté la dissolution de sels renfermant des ions Ca, comme par exemple $CaSO_4$, aura pour effet de diminuer la teneur en ions CO_3 et HCO_3. En somme, la teneur en CO_3 combiné se maintiendra en général entre certaines limites, une limite supérieure fonction de la tension du CO_2, pouvant légèrement monter suivant la teneur en ions autres que Ca, CO_3 et HCO_3, et une limite inférieure pouvant descendre suivant la teneur en ions Ca. D'une manière générale, le CO_3 combiné oscille entre 75 et 240 mg.

Si donc l'on fait exception des eaux acides issues des terrains cristallins ou gréseux purs, nous pouvons conclure, puisque les eaux sont à peu près saturées en carbonate de chaux, que si une eau a une concentration égale ou inférieure à la solubilité normale du $CaCO_3$, 300 à 400 mg, elle doit être presque exclusivement bicarbonatée calcique, le carbonate de chaux étant le premier sel à être dissous dans le sol agricole.

Le gypse est le sel suivant, le moins soluble et parmi les plus fréquents. En dissolvant le gypse, les eaux apportent des ions Ca qui par conséquent diminuent la teneur en CO_3 combiné de l'eau. Et il est remarquable de constater que les eaux très chargées en $CaSO_4$ ont toujours des teneurs en CO_3 combiné en dessous de la normale. La limite de solubilité en $CaSO_4$ peut être facilement atteinte en raison de l'abondance de ce sel, même naturellement être surélevée si les eaux dissolvent des sels renfermant des ions autres que Ca et SO_4. Les eaux chargées en NaCl peuvent ainsi avoir des teneurs en $CaSO_4$, beaucoup plus grandes. La solubilité du $CaSO_4 2H_2O$ (comptée en $CaSO_4$), de 2 gr, 1 à 20° pour une concentration nulle de NaCl, croit jusqu'à 7 gr, 30 pour une concentration de NaCl = 146 gr 2, puis décroit pour des concentrations supérieures. La solubilité pourrait être diminuée par une nouvelle dissolution de nouveaux sels de calcium. Mais parmi ceux-ci, le $CaCl_2$ n'existe pour. ainsi dire pas dans les roches et le $CaCO_3$ avait déjà atteint sa concentration maxima Mais un apport de sulfates, tels que Na_2SO_4 ou $MgSO_4$, peut amener le dépassement du produit de solubilité de $CaSO_4$ et par conséquent provoquer une précipitation de ce dernier sel. De tels précipités de sulfate de calcium ont été observés dans les roches. La dissolution du sulfate de magnésie est certaine dans les terrains gypsifères, car on constate que les eaux souterraines, à grandes valeurs de SO_4, ont des teneurs toujours très élevées en Mg, ce qui entraîne une augmentation des rapports Mg /Ca et Na /Ca.

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(1) H. SCHOELLER. — L'influence du climat ... p. 284, loc. cit.

Le sel le plus commun des eaux souterraines et possédant une solubilité supérieure aux précédentes, est le NaCl. Un long parcours, un temps de contact très grand, une grande surface de contact augmentent la teneur en NaCl de l'eau, mais sans amener la saturation qui n'a lieu que tout à fait exceptionnellement, la concentration en NaCl se mettant seulement en équilibre avec la concentration en NaCl du terrain. Pour qu'il y ait des valeurs voisines de la saturation, il faut que le terrain aquifère soit salifère. Ce n'est donc qu'exceptionnellement que l'on observe des valeurs voisines de la saturation.

En conclusion, on voit que plus une eau se charge en sels, mais au-dessus d'une certaine concentration correspondant à la saturation du sulfate de chaux, plus le rapport SO_4/Cl doit diminuer et plus les rapports Na/Ca, Na/Mg et même aussi le rapport Mg/Ca doivent augmenter.

La concentration par évaporation suit à peu près les mêmes règles que précédemment; et les variations des rapports des divers radicaux se font suivant le même ordre. Mais ici, il peut y avoir des précipitations plus nombreuses. La première précipitation dans le sol est celle du $CaCO_3$, sous forme de concrétions calcaires, observables dans les régions tempérées, et de tuf dans les régions steppiques ou prédésertiques; la seconde est celle du gypse dans les régions désertiques. Puis viennent les précipitations de sels de soude, carbonate de soude ou sulfate de soude, suivant la teneur en radicaux CO_3 et SO_4 de l'eau.

De tout ce qui précède, il résulte, en admettant que le CO_3 combiné ne dépasse pas 300 mg, qu'on doit généralement d'abord avoir dans les eaux (1) :

$$r\,CO_3 > r\,Cl \text{ ou } r\,SO_4 \quad (2)$$

et au-dessus d'une certaine concentration totale de sels dissous qui est généralement approximativement de 60 milliéquivalents :

$$r\,Cl \text{ ou } r\,SO_4 > r\,CO_3$$

et au-dessus d'une concentration encore plus élevée, généralement voisine de 180 milliéquivalents :

$$r\,Cl > r\,SO_4 > r\,CO_3$$

si les eaux n'ont pas dissous de sels de magnésie. Si évidemment il y a des sels de magnésie, SO_4 peut prendre des valeurs plus élevées. Nous avons constaté que c'est pratiquement au-dessus d'un concentration supérieure à 290 milliéquivalents que l'on a toujours $r\,Cl > r\,SO_4 > r\,CO_3$.

On voit donc que, si pour une raison quelconque une eau souterraine se concentre, que ce soit par évaporation d'une nappe arrivant tout près de la surface du sol, que ce soit dans le sol même de la zone d'alimentation, que ce soit par enrichissement progressif par circulation très lente, ou par suite de très longs cheminements où les questions de temps et de surface entrent en jeu, elle se transforme complètement, tendant toujours vers une eau de formule :

$$r\,Cl > r\,SO_4 > r\,CO_3 — r\,Na > r\,Mg > r\,Ca$$

et qui se rapproche de celle de l'eau de mer.

Nous donnons, ci-dessous, quelques exemples d'eaux très voisines de leur point de saturation, soit en NaCl, soit en $CaSO_4$.

	1	2	3	4	5	6
Ca	1071	6140	894	314	2799	2756
Mg	1974	2397	722	310	9307	4970
K	2155		1185	1004		
Na	88897	106160	96724	116598	51600	38456
Cl	137788	170500	150051	180365	106250	75000
Br	57		127	129		
SO^e	11981	1120	5241	2659	7544	8034
$CO_3+CO_3H/2$	162	60	205		138	99
résidu sec	245300	286000	256204	307790	177740	130960
$[SO_4]$ $[Ca]$	333×10^{-5}	179×10^{-5}	122×10^{-5}	22×10^{-5}	556×10^{-5}	574×10^{-5}
[Na] [Cl]	15,0	22,2	17,8	25,7		

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(1) H. Schoeller. — Sur la concentration des sels dissous dans les eaux souterraines. C.R. du Congrès d'Erfoud, 1934, du Comité d'Etude des Eaux souterraines, p. 46-54, Rabat, 1934.

(2) *r* signifie que les radicaux sont exprimés en milliéquivalents.

Eaux concentrées par dissolution : 1) Puits de l'oued Melah de la Saline de Lorbeus, Tunisie ; 2) Source salée, à 1 km au sud de Douira, près de Aïn Rehlal, Tunisie ; 3) Source Bayaa de Salies de Béarn, France, d'après Jacquot et Willm.

Eaux concentrées par évaporation: 5) Sebka à l'W de Kredma el Kebira, Tunisie ; 6) Sebka Kredma el Kebira, Tunisie.

Les analyses sont exprimées en milligrammes par litre.

Il existe des nappes d'eau à l'intérieur desquelles il n'y a aucune circulation, soit qu'il y ait égalité de pression hydrostatique sur tout le pourtour de la nappe, soit que la nappe soit complètement fermée. De même des portions de nappe peuvent avoir de l'eau stagnante, la circulation de l'eau ne se faisant que dans une partie de la nappe. Dans de telles *eaux de stagnation*, la concentration des sels a pu se faire totalement, n'ayant pas été limitée par le temps. L'eau doit donc être en équilibre ou sensiblement en équilibre avec la roche. Il suffit d'ailleurs d'une très petite quantité de sels solubles dans la roche et d'une très petite quantité d'éléments chimiquement attaqués par l'eau, pour donner une concentration élevée à cette eau.

Soient par exemple une roche de densité, d, contenant n % en poids de NaCl, m, la porosité. Il y a donc dans un mètre cube de roche, $1000\ d\ \mathrm{n}\ (1 - m)$ kg de NaCl, et dans l'eau d'imbibition, $1000\ d\ n\ (1 - m)$ gr de sels par litre, en admettant qu'il y ait équilibre de telle façon que la concentration dans l'eau soit la même que dans la roche, naturellement à volume égal. Si donc une roche ayant une porosité de 0,20 et une densité réelle de 2,65, renferme seulement 2 pour 1000 de NaCl, l'eau en équilibre avec la roche renfermerait 4 g, 240 de NaCl par litre. Les argiles peuvent renfermer 1 à 2 % de NaCl. Avec $d = 2{,}2$, $m = 0{,}40$, le calcul donne pour l'eau une concentration de 13,2 à 26 g, 4 de NaCl par litre.

Ces proportions correspondent à celles des roches d'affleurement déjà lessivées. Par conséquent les eaux profondes des terrains non lessivés peuvent avoir des concentrations de sels beaucoup plus élevées. Or la plupart du temps, la grande teneur en NaCl des eaux souterraines est choisi comme critère de la reconnaissance des *eaux connées*, c'est-à-dire des eaux emprisonnées par les sédiments, lors de leur précipitation et devenant par la suite fossiles. En réalité, comme on le voit, ce ne peut en être un, puisqu'il est commun à d'autres eaux. Une grande teneur en NaCl ne désigne pas forcément une eau de mer fossile, mais peut tout simplement signifier une concentration par évaporation ou seulement par dissolution, dans ce dernier cas souvent par des eaux souterraines stagnantes.

D'autres modifications non moins importantes peuvent avoir lieu dans les nappes d'eau souterraines. Nons ne faisons que les mentionner, car elles ont déja fait l'objet de nombreux travaux. Ce sont la réduction des sulfates et les échanges de bases.

La réduction des sulfates, accompagnée ou non de H_2S, suivant qu'elle est actuelle ou ancienne, amène naturellement un abaissement souvent considérable de la teneur en SO_4 qui peut même devenir insignifiante ou nulle. Mais ce qu'il y a lieu de remarquer, c'est qu'à cette diminution de SO_4, est corrélative une *augmentation souvent très notable de la teneur en* CO_3 *combiné*, ce sur quoi l'on n'a jamais insisté et qui pourtant est assez *caractéristique*. Les eaux ayant subi une réduction de sulfates, ont presque toujours une teneur en CO_3 combiné très au-dessus de la normale. En effet les réactions relatives à la réduction des sulfates peuvent être résumées de la manière suivante :

$$CaSO_4 + 2C \ \text{(réducteur)} = CaS + 2\ CO_2$$
$$CaS + 2\ CO_2 = Ca(HCO_3)_2 + H_2S$$

réactions qui ont lieu lorsque des eaux sulfatées sont en contact, en milieu privé d'oxygène, avec des matières hydrocarbonées, telles que des pétroles, des bitumes, des matières organiques décomposées et qui provoquent la formation de nouveaux bicarbonates.

En résumé, dans une eau ayant subi une réduction des sulfates, le rapport SO_4/Cl diminue et la valeur absolue du CO_3 combiné augmente.

Si la réduction des sulfates modifie les rapports entre les radicaux acides, les échanges de bases, eux, modifient les rapports entre les radicaux basiques. On sait en effet que les shales, les argiles, les roches glauconieuses sont susceptibles d'échanger leurs bases avec les eaux souterraines avec lesquelles elles sont en contact (1).

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(1) Renick Coleman-Base exchange in ground water by silicates, illustrated in Montana, U.S. Geological Survey Water Supply Paper, 520-d, 1924.

Mc Kenzie Taylor- The bearing of base exchange on the genesis of petroleum-Journ. of the Institut of Petroleum technologists, XIV, 835-840, 1928.

Id — The replaceable bases in the shales and clays overlying petroleums strata,Ibid., XV, 207-210, 1929.

Case L.C. — Base replacement studies of Oklahoma shales. Critiques of Taylor hypothesis. *Bul. of the americ. Assoc. of Petroleum geologists*, XVIII n° 1, 66-79, 1933.

Jung J. — Les théories de Mc Kenzie Taylor sur les échanges de bases dans les argiles et la géologie du pétrole. Ann. Office Nat. combustibles liquides, n° 2, 291-300, 1933.

Schoeller H. — Les échanges de bases dans les eaux souterraines. Trois exemples en Tunisie, *Bull. Soc. géologique de France*, (5), IV, 389-420, 1934.

Il y a des argiles sodiques cèdant leur sodium, des argiles calciques cèdant leur calcium :

$$\text{argile } Na_2 + (CO_3H)_2\,Ca \rightleftarrows \text{argile } Ca + 2\,NaHCO_3$$
$$\text{argile } Na_2 + CaSO_4 \rightleftarrows \text{argile } Ca + Na_2SO_4$$

qu'il serait plus correct d'écrire :

$$\text{argile } Na_2 + Ca^{++} \rightleftarrows \text{argile } Ca + Ca^{++}$$

et il y a également des échanges d'ions Mg

$$\text{argile } Na_2 + Mg^{++} \rightleftarrows \text{argile } Mg + 2\,Na^{+}$$
$$\text{argile } Ca + Mg^{++} \rightleftarrows \text{argile } Mg + Ca^{++}$$

Les rapports Mg/Ca, Na/Ca, Mg/Na des eaux se trouvent ainsi modifiés.

Mais ce sur quoi il faut surtout insister, c'est que dans cet échange de bases, il y a lieu de considérer :

1° la quantité absolue

2° La quantité relative des ions échangés.

La quantité absolue dans une eau est d'autant plus grande que l'eau était primitivement plus chargée en éléments échangeables. La quantité relative est d'autant plus grande, c'est-à-dire l'échange est d'autant plus complet en un temps donné, que la concentration est faible, et dans un même bassin aquifère que le temps de contact est grand. Plus la circulation est lente, plus le trajet est long dans une zone aquifère, plus l'échange est complet.

Aussi les eaux ayant subi une forte concentration par dissolution, montrent-elles un grand échange de bases.

De ce qui précède, il résulte qu'il y a lieu d'attacher une réelle importance à certaines valeurs absolues ou à certains rapports, lorsque l'on veut soit caractériser une eau, soit la comparer avec une autre.

D'une manière générale, les divers éléments chimiques ne varient, dans l'eau d'une même nappe ou dans l'eau de nappes provenant d'un même terrain, qu'entre certaines valeurs absolues, ou plutôt ont un maximum de fréquence entre ces valeurs, alors que pour d'autres nappes ou d'autres terrains, le maximum de fréquence est autre. (fig. 1). La concentration, soit la somme des divers éléments, suit évidemment la même règle. Par contre le CO_3 combiné reste d'une manière générale, comme nous l'avons vu, à peu près le même, quelle que soit la nature du terrain aquifère ou la nappe. Cependant les eaux ayant été en contact avec des matières carbonées ou des hydrocarbures qui par oxydation, donnent du gaz carbonique, de même que les eaux ayant subi une réduction de sulfates, ainsi que les eaux recevant du gaz carbonique profond d'origine métamorphique ou juvénile, peuvent être, plus que d'ordinaire chargées en CO_3 combiné. Par contre les eaux ayant été en contact avec l'air ont une teneur plus faible.

La teneur en H_2S est aussi une caractéristique ,signifiant une réduction de sulfate. L'absence d'oxygène, la présence d'NH_4 ou de NO_2 indiquent un milieu réducteur et peuvent caractériser une nappe.

Mais les éléments qui paraissent les plus constants, dans une même nappe ou dans un même terrain, sont les rapports caractéristiques rMg/rCa et rSO_4/rCl qui permettent de comparer réellement les eaux entre elles et d'identifier une nappe ou une portion de nappe (fig. 1). Ils ne semblent pas affectés par la concentration de l'eau lorsque celle-ci ne provoque pas le dépassement du produit de solubilité $[SO_4]$ [Ca]. Mais lorsque ce produit se trouve dépassé, rMg/rCa augmente avec la concentration, comme nous l'avons indiqué plus haut, et naturellement le rapport rSO_4/rCl diminue. Ces deux rapports sont donc, je le répète, parmi les plus importants à considérer, en particulier lorsque l'on compare des eaux entre elles. Enfin l'indice d'échange de bases $r\frac{Cl - (Na + K)}{Cl}$ ou $r\frac{Cl - (Na + K)}{SO_4 + CO_3}$, c'est-à-dire le rapport du nombre d'ions Na ou Ca et Mg échangés au nombre d'ions Na ou Ca et Mg primitivement existants, peut avoir une certaine valeur caractéristique.

Nous noterons toutefois que lorsque les eaux n'ont qu'une concentration relativement faible, moins de 0,5 milliéquivalents en Cl et SO_4, les rapports SO_4/Cl sont moins constants. On pourrait en dire autant des autres rapports caractéristiques. En effet une faible augmentation de la valeur absolue de l'un ou l'autre de ces éléments, entraine immédiatement une variation considérable de ces rapports.

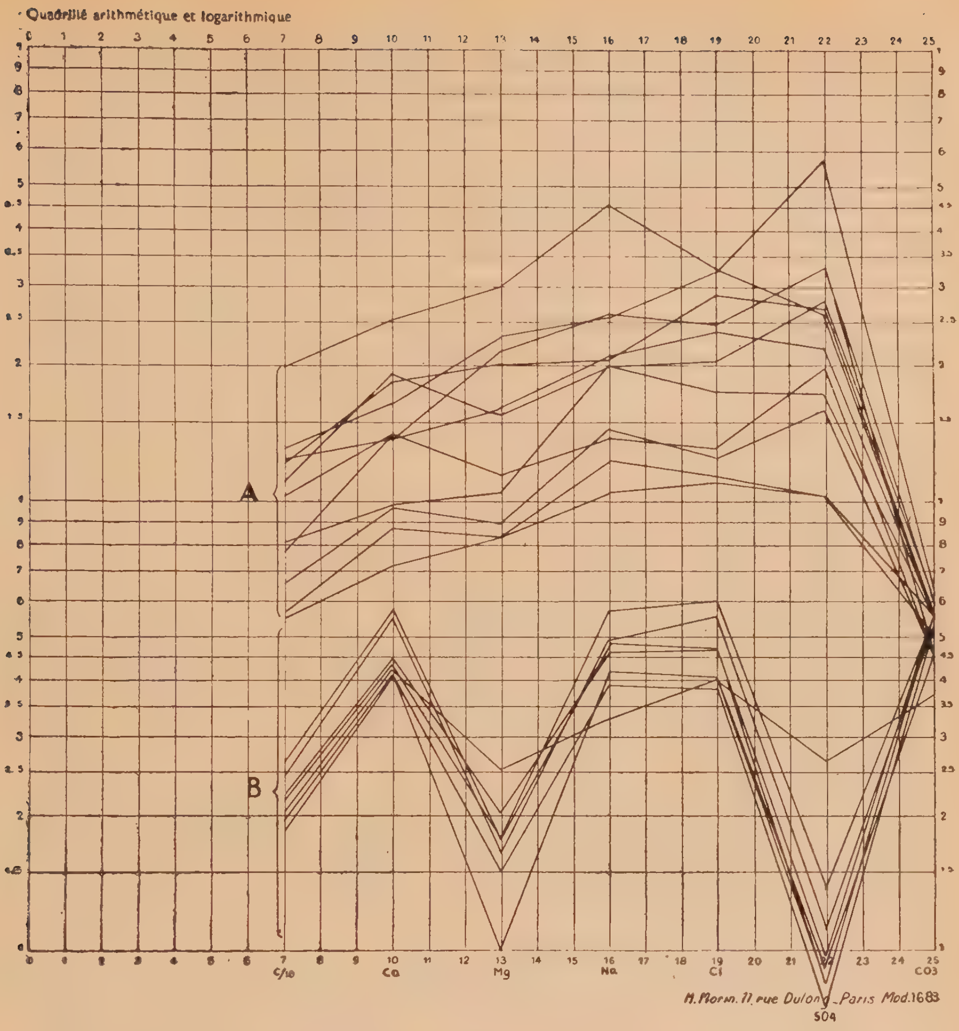

Fig. 1. — Composition chimique de l'eau de deux napes différentes
A : Nappe du Vindobonien d'Hadjeb el Aïoun, Tunisie
B : Nappe des dunes anciennes; quaternaires du Djebel Cherifate, près de Soliman, Tunisie. (L'une de ce eaux a une composition aberrante dont il faut rechercher la cause).
En ordonnée logarithmique : le nombre de milliéquivalents; C/10 = 1/10 de la concentration exprimées en milliéquivalents.

5) LES VARIATIONS DE LA COMPOSITION CHIMIQUE DE L'EAU DANS LES NAPPES SOUTERRAINES

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Les modifications précédemment examinées de la composition chimique des eaux, permettent de mieux comprendre, d'une part les différences de composition chimique de l'eau qui peuvent exister entre diverses nappes souterraines, et d'autre part les variations de la composition chimique de l'eau dans une même nappe.

A. — Variations de la composition chimique dans des nappes différentes

D'abord, je commencerai par formuler quelques règles, ou plutôt quelques remarques tirées de l'examen que j'ai fait d'un grand nombre d'analyses d'eau de nappes :

1° Les eaux issues d'une roche de même nature pétrographique, quel que soit l'âge du terrain, ou quelle que soit la nappe issue de ce terrain, peuvent avoir des caractères communs. Ainsi les eaux des calcaires auront généralement $r\,CO^3 > rSO^4$ ou rCl(1) et $rCa > r\,Mg$ ou rNa ; les eaux des grès et sables purs, de même que les eaux des roches cristallines, auront un résidu sec faible, un CO^3 combiné au-dessous de la normale ; les eaux en contact avec des éléments argileux ou marneux, un résidu sec élevé et une teneur plus grande en rCl et rNa, de telle sorte que l'on aura rCl ou $rSO^4 > rCO^3$ et $rNa > rMg$ ou rCa ; et les eaux issues des terrains gypseux, un résidu sec élevé et $rSO^4 > rCO^3$.

2° Mais toutes les eaux issues d'une roche de même nature pétrographique n'ont pas nécessairement exactement la même composition chimique, avec des rapports caractéristiques exactement les mêmes, car la diversité de l'alimentation de la nappe, la longueur du trajet des filets liquides et le climat apportent des modifications de composition chimique.

3° Les eaux issues de terrains de même nature pétrographique, de même âge, dans une même région, ont généralement des caractères communs. Et entre elles, il y a une parenté beaucoup plus grande qu'entre les eaux issues de terrains de même nature pétrographique mais d'âges différents et appartenant à des nappes différentes.

4° Mais les eaux de deux nappes différentes, situées dans ce même terrain, peuvent avoir des compositions chimiques différentes, même si ces deux nappes sont situées côte à côte. Plus les nappes sont éloignées les unes des autres, plus il peut y avoir de différences dans les compositions chimiques. En effet, l'alimentation des nappes peut ne pas être partout chimiquement la même. Les faciès eux-mêmes peuvent changer d'un point à un autre : Les différentes nappes peuvent avoir des trajets plus ou moins longs, amenant ainsi des variations dans les rapports caractéristiques. Dans certaines nappes, la circulation peut se faire de préférence, soit au sein même du terrain aquifère, soit au contact du mur, soit au contact du toit, amenant par là des modifications de la composition chimique.

Nous allons en donner quelques exemples :
Dans le premier, on a affaire à des nappes juxtaposées et ayant des compositions chimiques différentes. Il s'agit des nappes contenues dans le Quartenaire de la rive gauche de l'Oued el Hadjel, en Tunisie centrale, ou S.W. d'Hadjeb el Aïoun. Ce Quaternaire longe le flanc S.E. de l'anticlinal du Djebel ez Zaouia, formé de calcaire Cénomanien, entouré d'une ceinture de Pontien (2). Toutes ces couches sont plus

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(1) *r* placé devant le symbole ou un radical signifie qu'il s'agit de milliéquivalents.
(2) H. Schoeller, feuille Hadjeb-el-Aïoun de la carte hydrologique au 50.000 de la Tunisie, Tunis 1939.

ou moins prises en écharpe par la partie amont du Quaternaire. Ce sont donc tantôt des eaux des grès, des calcaires ou des dolomies du Cénomanien, tantôt des eaux des conglomérats pontiens, qui alimentent le Quaternaire. Aussi en cheminant d'une extrémité à l'autre de la bande du Quaternaire, rencontre-t-on des zones dans lesquelles les eaux ont des compositions chimiques quelque peu différentes. Dans la zone I, au S.W., rMg /rCa est voisin de I et rCl-rNa est positif. Dans la zone, 2, plus au N.E., rMgr /Ca est plus petit que 0,60 et rCl-rNa est négatif; tandis que dans la zone 3, au N.E., rMg /rCa est plus grand que 1,5 et rCl-rNa est négatif. Le rapport élevé rMg /rCa des bandes 1 et 3 est certainement dû à un apport d'eau provenant des dolomies du Cénomanien.

Mais si les eaux de ces trois zones diffèrent par le rapport rMg–rCa et l'indice d'échange de base, elles ont un même air de famille donné par le rapport $rSO^4 + rCl$.

Quaternaire de la rive gauche de l'Oued el Radjel

	Puits N°	Ca	Mg	Na	Cl	SO^4	CO^3	rs	rSO^4/rCl	rMgr /Ca	i.e.b.
I	2462	283	170	660	1109	990	147	3368	0,66	1,01	+ 8,1
	2463	355	210	781	1246	1244	180	4020	0,74	0,97	+ 3,2
	2465	208	135	570	896	747	159	2736	0,62	1,12	+ 1,8
	2466	346	264	1011	1565	1621	132	4949	0,77	1,29	+ 0,50
2	2469	485	323	1664	2462	252		7430	0,67	0,11	— 4,5
	2470	241	83	630	789	147		2790	0,66	0,57	— 10,9
	2471	584	132	2532	3470	2215	186	9140	0,47	0,36	— 2,41
	2472	292	103	809	1020	1196	135	3555	0,87	0,57	— 2,2
	2481	735	265	1945	2915	2687	168	8725	0,68	0,59	— 3,9
3	2487	180	180	685	975	991	123	3142	0,75	1,66	— 8,8

Les poids sont exprimés en milligrammes /litre.

i.e.b. est l'indice d'échange de base. S'il est positif, c'est $\frac{rCl — rNa}{rCl}$. S'il est négatif, c'est $\frac{rCl — 2Na}{rCO^3 + rSO^4}$. Autrement dit, c'est le rapport entre les ions Na ou Ca + Mg échangés et les ions Na ou Ca + Mg primitivement existants.

Les numéros des puits sont ceux du catalogue de l'Inventaire des Ressources hydrauliques de la Direction des Travaux Publics de Tunis.

Un autre exemple peut être fourni par la nappe du Quaternaire du Mornag, au S.E. de Tunis (1). Dans le même terrain cheminent, côte à côte, deux portions de nappe, séparées par une légère zone de drainage : la sous-zone de Créteville (C) et la sous-zone latérale (L) qui ont les caractères suivants :

	Résidu sec	rSO^4/rCl	rMg /Ca
C	< 1000	$< 0,30$	$< 0,30$
L	1000-3000	$> 0,30$	0,35-0,75

Ces deux sous-zones sont exactement dans le même Quaternaire. Mais tandis que la zone de Créteville descend du Kanghet el Hadjej, la sous-zone latérale reçoit les eaux du flanc du Djebel Ressas où affleure du Trias gypsifère et plus ou moins magnésien, contenant donc plus de sels solubles, d'où résidu sec plus élevé; plus de sulfates, d'où rSO^4/rCl plus grand, plus de sels de magnésie, d'où rMg /rCa plus grand.

Les exemples qui vont suivre, montrent que les eaux des nappes situées dans un même terrain, mais éloignées les unes des autres, ne sont point identiques, mais peuvent présenter des caractères communs :

Les calcaires éocènes de Tunisie sont aquifères, mais présentent de légères différences de faciès. Ainsi dans la région de Toukabeur et du Kef, ce sont des calcaires, souvent zoogènes, à Nummulites, en gros massifs; tandis que dans la région de Tunis, à Khledia et à l'Oued Bakbaka, ce sont des calcaires mieux lités, ainsi que dans la région de l'Oued el Kebir. Dans le Sud Tunisien, les calcaires sont moins bien développés. Comme on le voit dans le tableau ci-dessous, il y a des variations dans la concentration, les rapports rSO^4/rCl, rMg/rCa et dans l'indice d'échange de base. Dans la région de Toukabeur et du Kef, rSO^4/rCl est en général plus grand que dans la région de Khlédia, de l'Oued Bakbaka et de Bizerte, parce que les eaux peuvent descendre jusqu'à un mur imperméable formé de marnes plus ou moins gypsifères. Dans la région d'Hadjeb el Aïoun, le rapport est encore plus grand, car les marnes situées au contact des calcaires sont toujours très chargées en gypse.

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(1) Voir Fig. 1

		Concentration (Millivalences)	$\frac{rSO^4}{rCl}$	$\frac{rMg}{rCa}$	i.e.b.
1	Synclinal du Kef	11-14	0,59-1,40	0,20-0,25	+
	Synclinal de Toukabeur	14-19	0,49-0,71	0,28-0,42	+
2	Djebel Kebir à Bizerte	17-29	0,10-0,28	0,20-0,40	+
	Synclinal de Khlédia	22-36	0,20-0,55	0,29-0,52	+
	Synclinal de l'Oued Bakbaka....	17-19	0,33-0,41	0,18-0,30	±
3	Synclinal de l'Oued el Kébir	13-27	0,56-1,2	0,26-0,34	+
4	Hadjeb el Aïoun	42-70	0,63-2,1	0,75-1,2	±
	Sidi ben Gobrine	49-180	2,5 -3,6	0,73-1,9	+

Cependant, si les eaux de tous ces calcaires présentent des différences plus ou moins notables suivant les nappes d'où elles proviennent, chacune de ces nappes présente certains caractères propres. Et l'ensemble diffère également des eaux des autres terrains, comme par exemple des grès oligocènes par un rapport rMg/rCa plus petit.

En ce qui concerne les grès oligocènes de Tunisie, il y a une grande parenté entre les eaux du synclinal de Khlédia et celles du synclinal de l'Oued Bakbaka, synclinaux très voisins l'un de l'autre. Mais les eaux de l'Oued el Hadjel et de Bou Ficha provenant également de grès oligocènes ont un rapport rSO^4/rCl et des résidus secs plus grands, par suite de la proximité de formations séléniteuses.

Grès oligocènes de Tunisie	Concentration (Millivalences)	rSO^4/Cl	rMg/rCa	i.e.b.
Région d'Aïn Draham	2,9- 6	0,05-0,30	0,41-1,3	+
Synclinal Khlédia	11 - 41	0,23-0,58	0,47-0,80	±
Synclinal Oued Bakbaka	25 - 40	0,20-0,52	0,71-0,80	+
Bou Ficha	80 -160	0,45-0,52	0,90-1,30	—
Oued el Hadjel	90 -100	0,75-0,80	0,60	—

De même, dans les grès helvétiens du centre et du Sud de la Tunisie, le rapport rSO^4/rCl et parfois aussi le résidu sec sont plus élevés en raison de la proximité de formations gypseuses. Mais l'on remarquera que les eaux de ces grès, grès qui, surtout à la base, ressemblent souvent beaucoup à ceux de l'Oligocène, se distinguent des eaux des grès oligocènes par un rapport rSO^4/rCl plus élevé.

Grès helvétiens de Tunisie	Concentration (Millivalences)	rSO^4/rCl	rMg/rCa	i.e.b.
Synclinal de l'Oued Bakbaka	32- 55	0,30-0,71	0,57-1,0	—
M'Raissa, près Soliman	19- 32	0,20-0,45	0,50-0,80	—
Hadjeb el Aïoun	16- 55	1,6 -5,0	0,38-0,83	—
Sbeitla	21- 60	1,1 -2,3	0,40-1,00	—
Oued el Hadjel	80-150	0,75-1,1	0,63-0,72	—

Les exemples précédents peuvent servir aussi à illustrer le fait que les eaux de deux terrains de nature pétrographique différente, dans une même région, ont, à quelques exceptions près, une composition chimique différente. Ainsi, dans le synclinal de l'Oued Bakbaka, les eaux des calcaires éocènes ont une concentration plus faible et un rapport rMg /rCa plus petit que les eaux des grès oligocènes ou que les eaux des grès miocènes. Les mêmes faits peuvent se constater pour les terrains du synclinal de Khledia.

B. — Variations de la composition chimique de l'eau d'une même nappe

Les eaux d'une même nappe située dans un même terrain ont des caractères chimiques relativement constants, beaucoup plus constants que n'ont les eaux de deux nappes distinctes, même issues d'un même terrain, de même nature pétrographique et de même âge, même de la même assise géologique locale.

Mais la composition chimique de l'eau d'une même nappe peut varier d'amont en aval. A peu près toujours il se produit une concentration par dissolution, par conséquent une augmentation du résidu sec des eaux. D'une manière générale, le rapport rSO^4 /rCl diminue, la vitesse de dissolution étant plus grande pour les chlorures que pour les sulfates alcalino-terreux, surtout lorsque l'eau contient déjà une grande quantité de sulfates, ce qui est naturel puisque la vitesse de dissolution est proportionnelle au déficit de saturation, à peu près toujours plus grand pour les chlorures que pour les sulfates. Cependant, il est bien évident que si le terrain aquifère est riche en sulfates et relativement pauvre en chlorures, il peut y avoir d'abord augmentation du rapport rSO^4 /rCl, puis diminution lorsque SO^4 s'approche du point de saturation.

Le rapport rMg /rCa a tendance à augmenter. En effet, il n'y a plus apport de Ca par dissolution du $CaCO^3$, car les eaux en sont saturées dès l'origine. De plus, la dissolution du $Ca\,SO^4$ est moins rapide que celle du $Mg\,SO^4$ et du $Mg\,Cl^2$.

On peut également assister à des modifications de composition chimique à la suite de réductions de sulfates par les matières organiques, charbons de toute sorte, bitume, pétrole, etc... Enfin, l'indice d'échange de base a tendance à augmenter vers l'aval, lorsque cet échange est le fait de l'aquifère.

Naturellement, plus le trajet est long et plus la vitesse des filets liquides est petite, plus ces diverses variations s'accentuent vers l'aval.

Lorsque l'on a affaire à une nappe libre, soutenue, la surface piézométrique peut se rapprocher fortement de la surface du sol, à moins de 2 mètres, d'un mètre même. Il peut se produire alors, surtout dans les pays arides, une concentration par évaporation, par le jeu des eaux alternativement, ascendantes par remontée capillaire, et descendantes par infiltration. Le résidu sec augmente, le rapport rSO^4 /rCl diminue à la suite de la précipitation du sulfate de chaux, précipitation qui, conjointe à celle du carbonate de chaux, augmente le rapport rMg /rCa.

De plus, il est à remarquer que les terrains ne sont pas toujours homogènes dans toute l'étendue d'une même nappe, en particulier lorsqu'il s'agit de terrains clastiques comme le Quaternaire. Aussi conçoit-on que des filets liquides assez voisins, circulant dans des matérieux différents, puissent différemment dissoudre les substances et par là varier leur composition chimique.

Enfin, dans un terrain aquifère, l'eau circule le plus souvent en suivant des zones privilégiées, rarement d'une manière active dans toute la masse du terrain aquifère. Ceci permet une individualisation des filets liquides, les empêche de se mélanger et empêche la masse liquide d'avoir une composition parfaitement homogène.

Et à cela il faut ajouter que, dans le cas d'une nappe libre, il y a alimentation progressive d'amont en aval par toute la surface aquifère, ce qui peut entraîner des modifications par dilution.

Je vais donner ici deux exemples de modifications de composition chimique dans une même nappe.

Exemple de la nappe du Mornag (Tunisie) : Au Sud et au Sud-Est de Tunis s'étend une longue plaine de Quaternaire, la plaine du Mornag. Dans la partie orientale de cette plaine, entre l'Oued Miliane et l'Oued el Hamma, et le massif du Djebel Ressas et du Dj. Bou Kornine, une nappe d'eau circule alimentée en amont d'une part par la région du Kranget el Hadjag, et dautre part par la région du Ressas, et se déverse en Méditerranée, entre Hammam-Lif et l'embouchure de l'Oued Miliane.

Le Quaternaire est formé de limons et de sables, avec des matériaux plus grossiers, galets et sables grossiers en amont. La nappe qui s'y meut est libre, mais soutenue à la fois par l'Oued Miliane et la Méditerranée. Les couches isopiézométriques

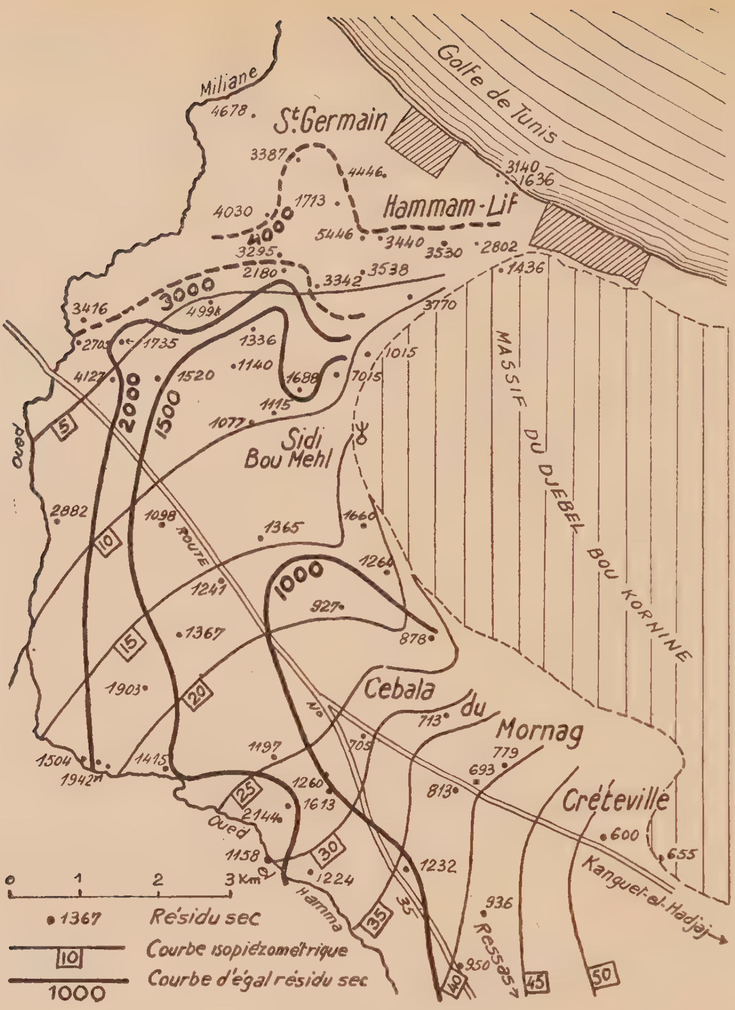

Fig. 1. — Nappe du Normag : le résidu sec.

montrent qu'elle est limitée au N.E. par une zone de drainage, située au pied des, contreforts du massif du Bou Kornine, et bien nette à l'Est de la Cebbala du Mornag jusqu'à Sidi Bou Mehl. L'Oued Miliane, constituant lui aussi une zone de drainage, la limite au S.W. et à l'Ouest.

Mais on constate que la nappe du Mornag peut se subdiviser en :

a) Une sous-nappe de Créteville, limitée à l'Est par la zone de drainage du bord oriental de la nappe du Mornag et à l'Ouest par une petite zone de drainage qui suit approximativement la route n° 35 de Tunis à Bou Ficha. Vers le Nord, elle s'étend un peu au Nord de la Cebbala du Mornag. Cette sous-nappe de Créteville descend directement du Kanghet el Hadjaj. Elle s'individualise en particulier par la plus grande vitesse de ses filets liquides, probablement par suite d'une plus grande alimentation, la suralimentation par l'agriculture n'y étant vraisemblablement pas étrangère.

b) Une sous-nappe latérale, limitée d'une part par la zone de drainage de la route 35 et d'autre part par l'Oued el Hamma. L'eau de cette nappe provient en partie du massif du Dj. Ressas. Des filets liquides de la nappe précédente peuvent d'ailleurs s'y ajouter.

c) Une sous-nappe aval, formée par la coalescence progressive des deux sous-

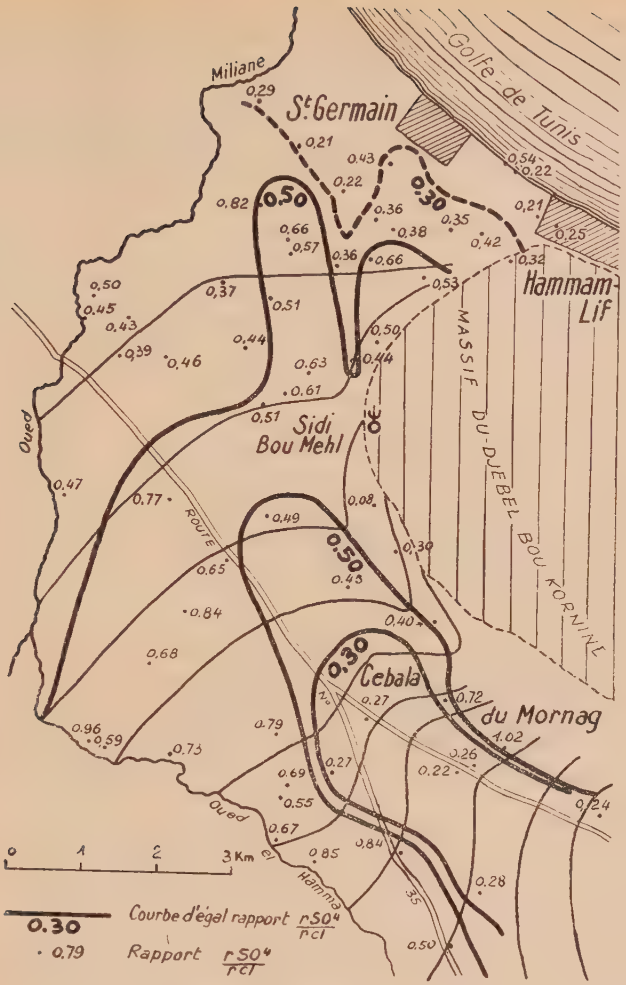

FIG. 2. — Nappe du Mornag : le rapport rCl/rSO^4.

zones précédentes, à partir de la Cebala du Mornag. Dans la sous-zone de Créteville, le résidu sec est relativement faible, inférieur à 1000 mg., ce qui va de pair avec une circulation plus grande de l'eau. rSO^4/rCl est inférieur à 0,30. C'est le rapport caractéristique des eaux de la nappe descendant du Kanghet el Hadjaj, où les terrains gypseux n'abondent pas particulièrement. Le rapport rMg/rCa est aussi relativement faible.

Dans la sous-zone latérale, l'eau provenant des flancs du Djebel Ressas où le Trias gypseux affleure, il est naturel que le résidu sec soit plus élevé et les rapports rSO^4/rCl et rMg/rCa plus grands, le gypse se dissolvant facilement, amenant donc plus de substances et en particulier des radicaux SO^4 et Mg. Ici encore on peut noter, comme nous l'avons précédemment vu, que deux nappes situées dans un même terrain peuvent présenter des eaux de composition chimique différente, même si ces nappes sont juxtaposées. On remarquera la coïncidence des courbes résidu sec = 1000, rSO/rCl = 0,50 et rMg/rCa = 0,30. avec la zone de drainage séparant les deux sous-zones. De part et d'autre de cette zone de drainage, les eaux sont donc bien différentes. Dans la sous-zone aval, formée par la confluence des deux sous-zones précédentes, mais dans laquelle les filets liquides peuvent garder leur individualité, le résidu sec s'accroît vers l'Oued Miliane et la mer. C'est un accroissement par dissolutions nouvelles dans le terrain aquifère. Le rapport rSO^4/rCl semble d'abord s'accroître légèrement, mais par la suite diminue nettement, surtout en direction de la mer, alors que le rapport rMg/rNa subit un accroissement très net vers l'aval. Ces deux modifications dues aux nouvelles dissolutions, confirment donc ce que nous avons précédemment dit.

Tout à fait au voisinage de la mer, en aval de la courbe rMg/rCa – 1,00, la nappe se trouve le plus souvent à moins de 2 mètres de profondeur. L'évaporation peut alors jouer son rôle, en accentuant les modifications chimiques dues aux dissolutions nouvelles.

Analyses des eaux des puits de la nappe du Mornag

Puits N°	H	Ca	Mg	Na	Cl	SO^4	CO^3	rs	pH	$\frac{rSO^2}{Cl}$	$\frac{rMg}{Ca}$	i.e.b.
Sous-zone de Créteville												
626	25.5	122	26	86	175	78	165	655	7.5	0.33	0.36	+23.9
622	19.0	118	29	65	160	52	174	600	7.2	0.24	0.41	+37.1
651	10.2	75	4	271	382	144	54	936		0.28	0.09	—21.2
620	10.15	147	30	70	190	97	156	693	7.1	0.26	0.34	+58.7
618	8.40	174	34	110	290	86	192	893	7.0	0.22	0.30	41.4
616	5.80	128	31	92	210	76	165	705	7.2	0.27	0.40	+32.4
681	4.9	246	39	151	410	235	174	1260	7.2	0.27	0.26	+43.2
760	8.3	186	35	97	275	166	162	927	7.1	0.45	0.31	+45.6
621	19.50	164	24	81	151	208	147	779	7.2	1.02	0.24	+17.2
631	12.70	143	26	80	160	155	144	713	7.3	0.72	0.30	+22.6
661	20.80	160	34	112	270	146	150	878	7.2	0.40	0.28	+35.9
Sous-zone latérale												
644	12.5	194	32	96	298	203	123	950		0.50	0.27	+50.3
643	9	232	38	155	310	354	135	1232		0.84	0.27	+22.6
678	5.5	197	54	167	295	340	168	1224	7.4	0.85	0.51	+12.6
657	5.7	414	94	200	735	547	150	2144	7.1	0.55	0.38	+58.0
676	8.35	284	62	198	470	438	156	1613	7.1	0.69	0.36	+35.0
674	6.4	215	49	130	320	312	112	1197	7.2	0.72	0.38	+37.2
Sous-zone aval												
687	9.1	172	54	255	395	390	147	1415	7.4	0.73	0.52	0
692	7.02	199	65	226	375	487	147	1504	7.3	0.96	0.54	+ 7.0
691	7.95	186	91	361	640	507	153	1942	7.2	0.59	0.81	+12.9
747	5.9	256	84	292	571	529	162	1903	7.6	0.68	0.55	+66.5
716	5.0	197	54	197	355	402	153	1367	7.2	0.84	0.46	+14.4
721	8.05	220	48	150	354	309	153	1241	7.2	0.65	0.36	+34.6
780	9.2	199	59	216	430	253	201	1365	7.2	0.49	0.49	+22.5
734	9.5	282	130	554	1065	691	144	2882	7.5	0.47	0.77	+19.7
729	8.6	180	50	132	282	292	155	1098	7.2	0.77	0.46	+27.7
861	8.55	186	55	121	320	222	168	1077	7.2	0.51	0.49	+41.6
864	6.9	192	61	122	305	251	177	1115	7.4	0.61	0.53	+38.3
883	6.2	265	85	221	505	421	189	1688	7.2	0.61	0.53	+32.5
874	8.20	174	55	157	361	213	171	1140	7.2	0.44	0.53	+32.9
877	5.0	190	56	208	425	292	153	1336	7.4	0.52	0.49	+24.5
854	8.2	163	68	295	495	307	183	1520	7.6	0.46	0.69	+8.03
852	7.4	175	74	349	610	356	159	1735	7.5	0.43	0.70	+11.7
847	7.6	463	201	722	1665	875	189	4127	7.6	0.39	0.72	+33.0
844	7.5	289	119	511	985	607	189	2705	7.2	0.45	0.69	+19.8
843	8.8	369	144	634	1210	811	234	3416	7.2	0.50	0.65	+19.1
868	7.3	678	265	715	2115	1053	159	4994	7.1	0.37	0.65	+47.7
903	6.35	539	174	411	1595	518	189	3342	7.1	0.26	0.54	+57.6
928	5.05	265	118	344	695	538	204	2180	7.2	0.57	0.74	+23.6
932	4.70	542	250	948	1765	1576	195	3295	7.4	0.66	0.77	+17.1
900	3.70	360	186	616	1160	1043	159	3538	7.6	0.66	0.86	+18.1
897	4.5	331	178	758	1395	964	129	3770	7.9	0.53	0.90	+18.2
38	4.30	453	179	688	1200	1328	186	4030	7.3	0.82	0.66	+11.6
30	3.20	497	312	1025	2294	1108	174	5446	7.7	0.36	1.05	+30.7
29	2.80	276	185	708	1353	692	195	3440	7.4	0.38	1.12	+19.3

28	2.0	273	157	795	1430	681	171	3530	7.3	0.35	0.96	+14.2
31	1.0	121	114	766	970	550	240	2802	7.9	0.42	1.57	—28
939	2.70	40	41	551	500	305	264	1713	8.3	0.22	1.71	—65.1
950	1.75	331	199	1010	1665	980	261	4466	7.6	0.43	1.00	+ 1.9
942	2.75	76	86	1086	1470	426	231	3387	8.4	0.21	1.88	—37.2
946	4.25	225	173	1256	1985	793	228	4678	7.8	0.29	1.28	+ 0.6

Les poids sont exprimés en milligrammes/litres— rs = résidu sec — i.e.b. = indice d'échange de bases — positif, c'est r $\frac{Cl\text{-}Na}{Cl}$ —négatif, c'est r $\frac{Cl - Na}{CO^3 + SO^4}$ — r placé devant un symbole et un radical signifie qu'il s'agit de milliéquivalents. — Les N° des puits sont ceux du catalogie du Bureau d'Inventaires des Ressources Hydrauliques de la Direction des Travaux Publics de la Régence de Tunis. — H = la profondeur, sous le sol, de la surface de l'eau du puits. Les puits peuvent être repérés sur les cartes ci-jointes à l'aide du résidu sec.

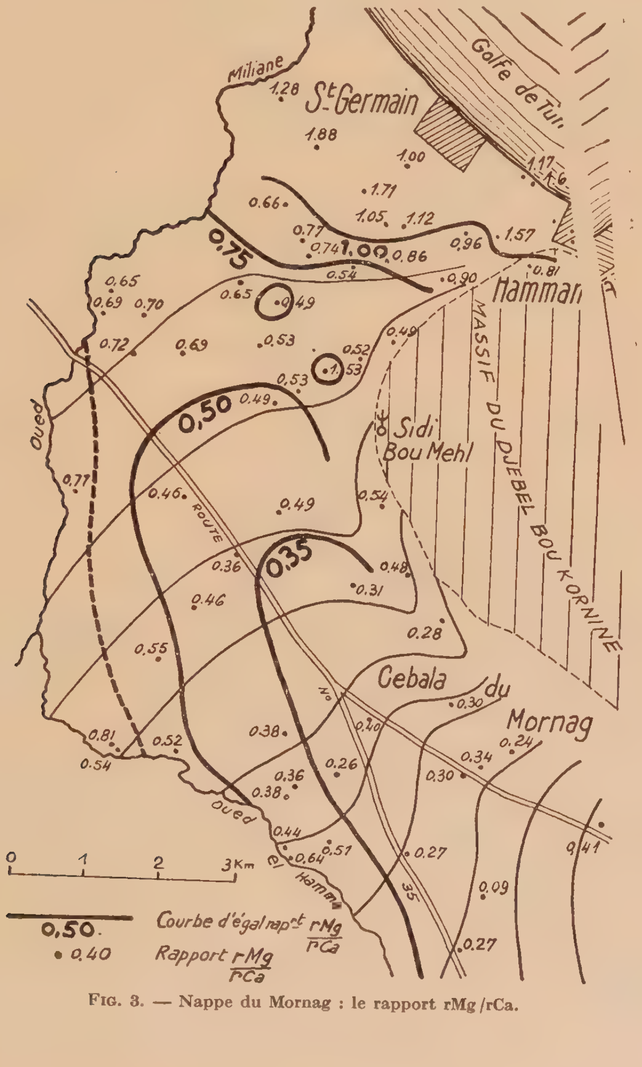

FIG. 3. — Nappe du Mornag : le rapport rMg/rCa.

Nappe de l'Oued el Fekka, plaine de Gamouda (Tunisie) : Le terrain aquifère de cette nappe située dans une grande plaine limitée par quelques massifs montagneux, est surtout formé de limons, de sables avec parfois des graviers.

La nappe est délimitée, grâce aux courbes isopiézométriques, par une zone de drainage au pied du Djebel Rakhmat et aboutissant au Bled Oum el Adam, et au S.E. par une autre zone de drainage qui suit à peu près le tracé du Sarg Diba (1).

L'alimentation de la nappe se fait surtout par infiltration de l'Oued Fekka, dans la partie amont de la plaine de Gamouda, Quant à l'écoulement de la nappe, il se dirige vers l'Oued en Neggada, au N.N.E. C'est l'écoulement majeur dont une partie est drainée par l'Oued en question. Mais il y a également un écoulement vers la Garaet el Akrech et l'Aïn Rebaou.

Examinons maintenant l'évolution de la composition chimique de la nappe, au cours de son trajet, de la zone d'alimentation à la zone aval.

Dans la zone d'alimentation, le résidu sec est relativement faible, inférieur à 2000 mg. Mais il augmente vers l'aval, vers la zone de drainage de l'Oued el Fekka, en particulier en direction du Bled Oum el Adam, où il atteint 5000 mg–litre en arrivant au bord de la zone d'évaporation. Dans la région, en amont de la courbe de résidu sec 5000, tous les puits ont plus de 2 mètres de profondeur, de telle sorte que l'accroissement du résidu sec ne peut être dû qu'à la concentration par dissolution.

Mais en aval de la courbe 5000, le résidu sec augmente très rapidement et peut

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(1) On consultera : H. Schoeller : « *Les échanges de bases dans les eaux souterraines vadoses. Trois exemples en Tunisie* »: Bull. Soc. Géologique de Fr. (5), 405, 1934.

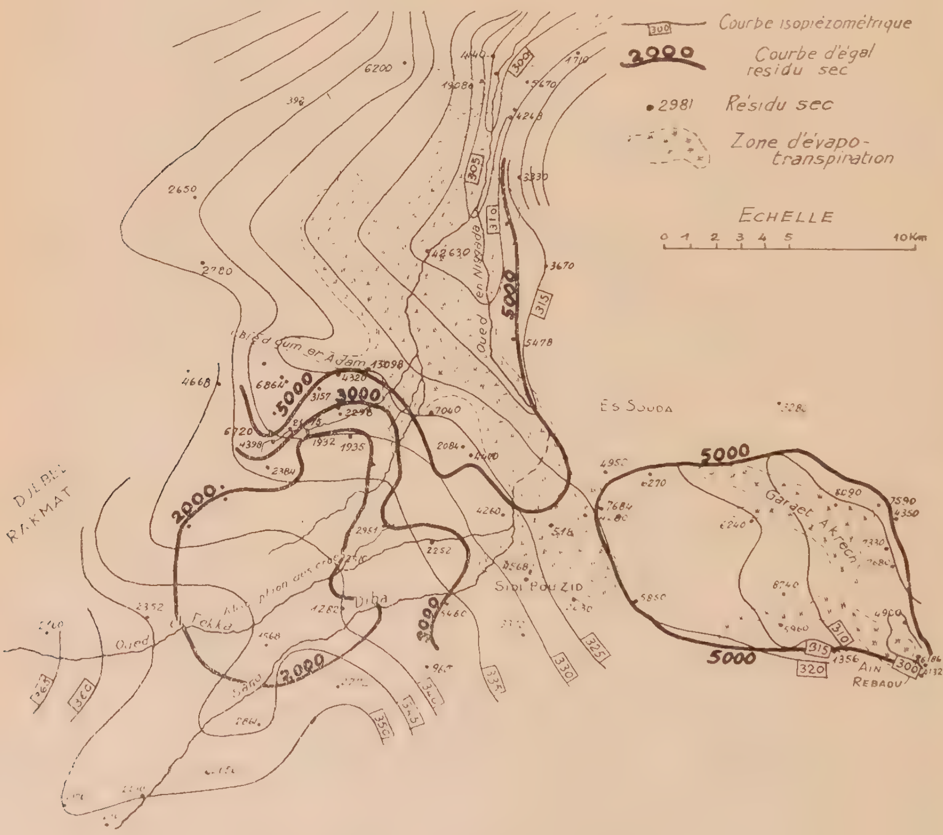

Fig. 4. — Nappe de l'oued el Fekka : le résidu sec.

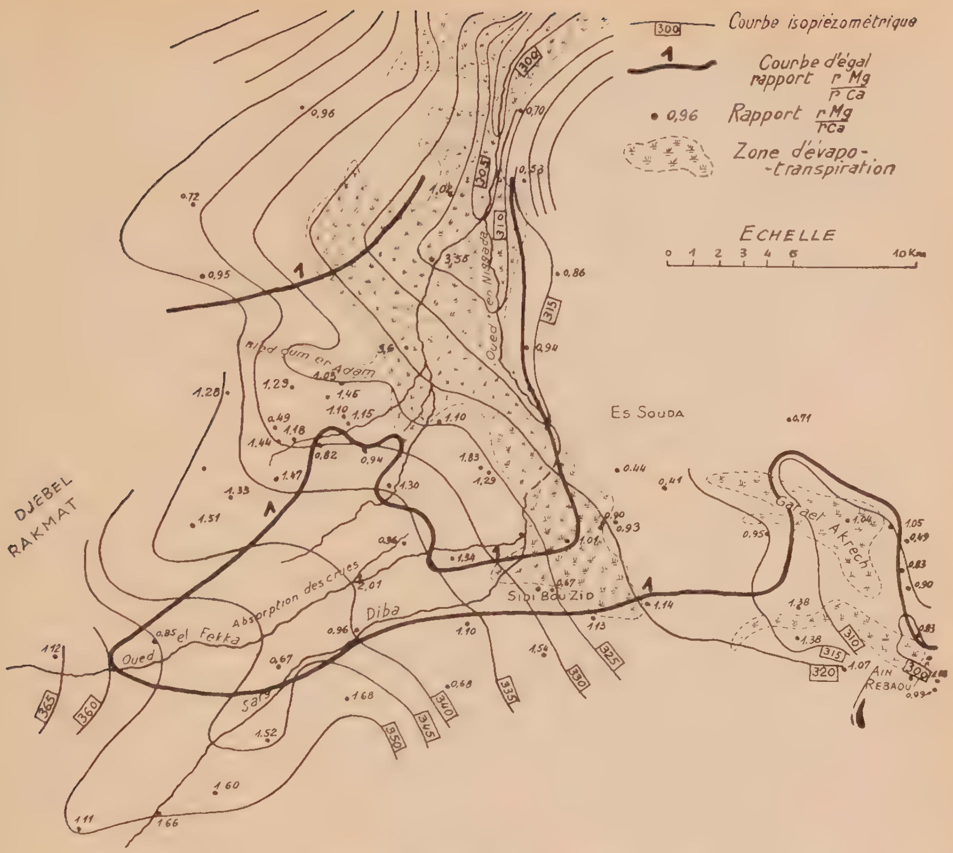

FIG. 5. — Nappe de l'oued el Fekka : le rapport rCl/rSO^4.

atteindre 42.630 mg/l. Cette augmentation est évidemment due à l'évaporation, dans la zone de drainage de l'Oued en Neggada, où l'eau est à bien moins de 2 mètres de profondeur et où le sol est couvert d'halophytes.

En direction de Sidi Bou Zid et d'Es Souda, l'accroissement du résidu sec commence plus tôt, mais il est plus lent ; ce qui peut être la conséquence du résidu des crues de l'Oued Fekka augmentant plus tôt le résidu sec des eaux, et des crues pénétrant à l'aval dans la nappe. De là les eaux de la nappe se dirigent vers la zone de drainage de l'Aïn Rebaou. Pendant ce trajet, leur concentration augmente par dissolution, dépasse 5000 mg, puis par évaporation dans la plaine salée où elle atteint plus de 8.000 mg.

Dans la zone d'alimentation, le rapport rSO^4/rCl est supérieur à 5. C'est le rapport caractéristique des eaux d'infiltration de l'Oued el Fekka. Ce rapport s'abaisse jusqu'à 3, d'une part vers la zone de drainage N.W. et vers la zone de drainage de l'Oued el Fekka où il peut descendre jusqu'à 1. Cette diminution du rapport rSO^4/rCl ne peut être due qu'à une augmentation de la teneur en chlorures, et cela par dissolution, puisque toute cette zone se trouve en dehors de la zone d'évaporation.

Par contre, dans la zone de l'Oued el Fekka où la nappe est à moins de deux mètres du sol, le rapport rSO^4/rCl tombe à moins de 1. Ici, c'est l'évaporation qui intervient surtout par précipitation de SO^4Ca et de SO^4Na^2 qui apparaît en efflorescence à la surface du sol. Les mêmes phénomènes se produisent vers la zone de drainage de l'Aïn Rebaou, diminution de rSO^4/rCl jusqu'à 1 par dissolution et jusqu'à des valeurs inférieures à 1 par évaporation. Mais ici la diminution du rapport est aussi partiellement due à l'arrivée d'eaux venant du Quaternaire du Sud de Sidi Bou Zid, ces eaux ayant un rapport $rSO^4/rCl < 3$.

	rSO^4/rCl > 5					5 > rSO^4/rCl > 3										3 > rSO^4/rCl > 1			< 1
Puits	62-11	94-3	92-8	92-13	92-16	92-11	40-9	93-3	92-1	400	92-9	92-19	92-17	62.10	92-6	92-3	92-7	91-24	93-10
Ca	230	567	206	190	318	217	191	364	364	211	141	164	171	166	534	551	201	485	735
Mg	93	124	116	229	162	174	166	226	274	104	110	131	155	96	319	604	139	293	1568
Na	125	208	251	287	190	236	618	750	600	264	340	248	172	118	1308	1560	311	808	12348
Cl	90	168	152	210	195	210	285	400	410	210	200	200	190	140	900	1400	355	1200	16000
SO^4	912	1707	1054	1440	1332	1269	1724	2410	2418	1153	1070	946	908	641	3734	4701	907	2298	11729
CO^3	108	207	154	150	150	120	162	150	171	123	120	129	132	111	216	252	210	75	180
rs	1568	2984	1935	2510	2352	2252	3157	4320	4260	1932	2010	1830	1740	1282	7040	9084	2158	5180	42630
dh	94	190	97	140	144	124	113	180	200	93	80	94	104	80	260	380	106	236	810
pH	7.2	7.8	9.7	7.4	7.2	7.4	8.1	8.0	7.5	7.6	8.0	7.5	7.4	7.3	7.6	7.8	7.6	7.4	8.0
rSO^4 rCl	7.44	7.44	5.15	5.07	5.04	4.49	4.46	4.43	4.39	4.07	3.57	3.50	3.50	3.38	3.21	2.49	1.89	1.41	0.54
rMg rCa	0.67	0.36	0.94	2.00	0.84	1.25	1.45	1.03	1.25	0.82	1.30	1.33	1.50	0.96	0.99	1.83	1.15	1.01	3.55
i.e.b.	—0.07	–0.10	–0.21	–0.19	–0.07	–0.14	–0.41	–0.38	–0.26	–0.35	–0.35	–0.21	–0.09	–0.07	–0.37	–0.27	–0.14	–0.03	–0.34
H.	22.80	0.20	4.50	14.80	30.50	5.80		0.0	2.80	10.35	6.65	18	23.60	13.4	5.00	3.90	3.80	2.80	0.00

rMg/rCa < 1

2 < rMg/rCa > 1

> 2

résidu sec > 5000

H est la profondeur de la surface de l'eau du puits, sous le niveau du sol. Les poids sont exprimés en milligrammes par litre — rs = résidu sec — dh = degré hydrométrique — i.e.b. indice d'échange de base (cf. p. 120).

Les différents puits peuvent être reprérés sur les cartes, grâce au résidu sec.

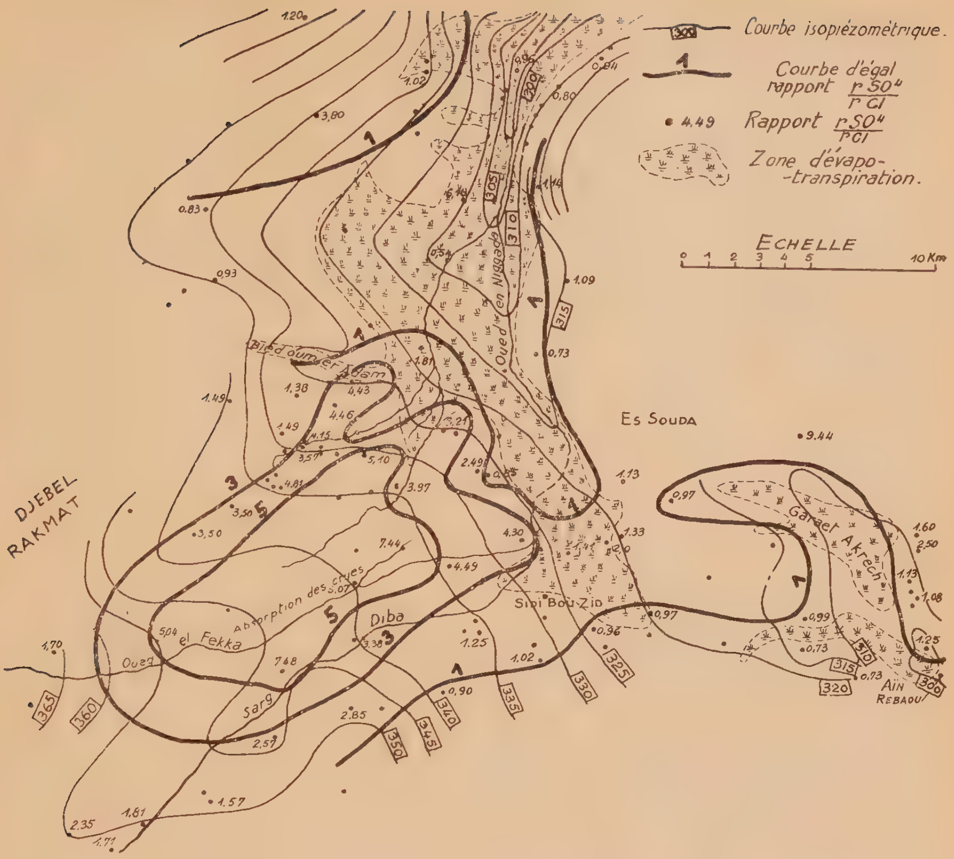

Fig. 6. — Nappe de l'oued el Fekka : le rapport rMg/rCa.

Dans la zone d'alimentation, le rapport rMg/rCa est plus petit que 1. En aval, il croît par dissolution, le Mg pouvant être apporté par les chlorures, surtout par les sulfates nouvellement dissous. Dans la zone de l'Oued en Neggada, il peut même atteindre 3, ce que l'on peut interpréter comme étant dû à une précipitation des sels de Ca, SO^4Ca, CO^3Ca, par concentration par évaporation.

Les eaux de la zone d'alimentation n'ont qu'un faible indice d'échange de base inférieur à + 0.10. Très rapidement vers l'aval, cet indice passe à — 0.10, dépasse — 0.20 et même — 0.30 dans la zone d'évaporation. Cet échange dû au contact avec les terrains aquifères est, il faut le noter, moindre le long du trajet des cours d'eau, par suite de l'alimentation directe de la nappe dans ces zones.

La nappe des sables verts de l'Albien du bassin de Paris.

Des modifications chimiques semblables à celle des nappes libres ont lieu dans les nappes captives. Telles sont, parmi les plus remarquables, la concentration, les échanges de base et la réduction des sulfates, qui peuvent être en particulier suivies dans la nappe des sables verts de l'Albien du Bassin de Paris.

Cette nappe est alimentée, au moins pro parte, par les affleurements qui s'étendent de Hirson à Vierzon, en passant à l'Est de Vouziers, Sainte-Menehould, à Saint-Dizier et au S.E. de Troyes. Dans la Haute-Marne, le niveau piézométrique est entre 120 et 140 mètres. Il est plus élevé au S.W., dans l'Yonne, où il est à plus de 200 mètres. Dans la partie captive de la nappe, il est inférieur à ces valeurs. A l'origine, il était à 125 mètres à Paris. Dans la Seine-Inférieure, l'Eure, la Somme, les niveaux piézométriques sont beaucoup plus bas, sauf autour de l'anticlinal du Pays de Bray qui fonctionne comme zone de dispersion. Par conséquent, l'eau circule certainement des affleurements vers la Manche et la nappe doit se déverser dans la Manche, car son niveau piézométrique est supérieur à celui de la mer. Il n'y a donc pas possibilité d'entrée d'eau de mer dans la nappe.

J. Delécourt (1) a divisé la nappe des sables verts en plusieurs zones chimiques :

1° — Une zone de dessalure, avec $rMg + rCa > 1$.

2° — Une zone des échanges de base, avec $rMg + rCa > 1$ mais voisin de ce chiffre. C'est une zone limite.

3° — Une zone de salure, avec $rMg + rCa < 1$. et dont les eaux sont plus chargées en sels que celles des zones précédentes.

4° — Une zone de sursalure, avec $rMg + rCa < 1$ et eaux très chargées en sels.

Pour chacune de ces zones, Delecourt a donné la composition chimique des eaux indiquée dans le tableau suivant. J'y ai ajouté les caractéristiques rSO^4/rCl, rMg/rCa, $\frac{rCl - Na}{SO^4 + CO^3}$, $\frac{rCl - Na}{Cl}$.

Forages	rCa	rMg	rNa	rCl	rSO⁴	rCO³	C	ieb	rSO⁴ / Cl	rMg / Ca
Zone des Affleurements										
I-St-Dizier	1.77	0.18	0.28	0.24	0.39	1.6	4.46	—0.02	1.62	0.10
II-Vierzon	1.67	0.21	0.20	0.15	0.20	(1.75)	4.16	—0.02	1.33	0.13
Moyenne de I et II	1.72	0.19	0.24	0.19	0.29	1.67	4.30	—0.02	1.52	0.11
Zone parisienne de dessalure										
V. Butte aux CaillesParis	1.39	0.66	(0.71)	0.20	0.26	2.30	5.52	—0.20	1.30	0.47
VI. GrenelleParis	1.32	0.71	(0.79)	0.23	0.21	2.38	5.64	—0.22	0.91	0.54
VII. Passy Paris	1.39	0.71	(0.73)	0.23	0.24	2.36	5.66	—0.19	1.04	0.51
VIII. Nanterre	1.50	0.62	0.79	0.25	0.24	2.42	5.82	—0.20	0.96	0.41
IX. Maison-Laffite	1.35	0.70	(0.94)	0.23	0.20	2.56	5.98	—0.26	0.87	0.52
X. Rue Blomet(P)	1.39	0.75	0.88	0.20	0.25	2.57	6.04	—0.24	1.25	0.54
Moyenne	1.39	0.69	0.81	0.22	0.23	2.43	5.77	—0.22	1.04	0.50
Zone des échanges de bases										
XI. Villemomble	1.34	0.62	1.35	0.31	0.27	2.72	6.61	—0.34	0.87	0.46
XII. P antin	1.41	0.77	1.49	0.27	0.20	3.25	7.39	—0.35	0.74	0.55
XIII. Plaine St-Denis	1.32	0.58	1.44	0.30	0.24	2.80	6.68	—0.37	0.80	0.44
Moyenne	1.36	0.66	1.42	0.29	0.24	2.92	6.89	—0.36	0.83	0.48
Zone de salure										
XVI. L'IsleAdam	0.55		4.42	0.60	0.37	4.00	9.94	—0.87	0.62	
XVIII. Bucy-le Long	0.33	0.61	8.98	0.70	3.21	6.00	19.82	—0.90	4.59	0.58
Zone de sursalure										
XIX. Eu	0.80	0.59	22.98	16.88	1.59	5.90	48.74	—0.82	0.09	0.74
XX. Le Crotoy	2.62	3.12	74.79	66.00	7.93	6.60	161.06	—0.60	0.12	1.19
XXI. Amiens	3.49	4.19	89.82	46.00	41.10	10.40	195.00	—0.85	0.89	1.20

i.e.b. négatif $= \frac{rCl\text{-}Na}{SO^4 + CO^3}$ i.e.b. positif $= \frac{rCl\text{-}Na}{Cl}$

C = Concentrations

Les valeurs sont données en milliéquivalents.

Examinons les modifications chimiques qui ont lieu d'une zone à l'autre. Elles peuvent être représentées par les augmentations ou diminutions de la moyenne de chaque élément chimique, ce que nous donnons dans le tableau suivant, dans lequel Δ *a* = différence entre zone des affleurements et zone de dessalure. Δ *b* = différence entre zone de dessalure et zone des échanges, Δ*c* = différence entre zone des échanges et zone de salure, Δ *d* = différence entre zone de salure et zone de sursalure.

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(1) J. Delecourt : « Les eaux artésiennes salines du Bassin de Paris, de la Basse et de la Moyenne Belgique » *Bull. Soc. belge de Géologie*, XLVII, 501, 1937.

Δ	rCa	rMg	rNa	rCl	rSO^4	rCO^3	C
Δ*a*	—0.33	+0.50	+ 0.57	+ 0.03	— 0.06	+0.76	+ 1.47
Δ*b*	—0.03	—0.03	+ 0.61	+ 0.07	+ 0.01	+0.49	+ 1.12
Δ*c*	—1.03	—0.05	+ 7.56	+ 0.4	+ 2.97	+3.08	+ 12.93
Δ*d*	+3.16	+3.58	+80.84	+45.30	+37.89	+4.4	+175.18

De la zone des affleurements à la zone dite de dessalure, et de la zone de dessalure à la zone des échanges de base, il y a déjà quelques modifications très sensibles, en général progressives, augmentation faible de la concentration du Cl, apport de CO^3 combiné peut-être dû à la transformation des carbonates en bicarbonates par suite de la cession d'ions Na contre des ions H et Ca de l'eau, échange de base notable qui passe de la première à la seconde zone de —0.02 à déjà —0.20, et de la seconde à la troisième zone de —0.22 à —0.36. Mais de la zone des échanges de base à la zone de salure (cf. Δ*c*), l'augmentation de la concentration est déjà considérable. Il y a un enrichissement, spécialement en SO^4, Na et Cl par dissolution, et augmentation de CO^3 combiné. Mais l'origine de ce CO^3 ne peut s'expliquer comme précédemment. Peut-être ce CO^3 est-il d'origine profonde. L'accroissement du Na est non seulement dû à de nouvelles dissolutions, mais aussi à l'échange contre des ions Ca et Mg de l'eau. Aussi l'indice d'échange de base passe-t-il à —0.90.

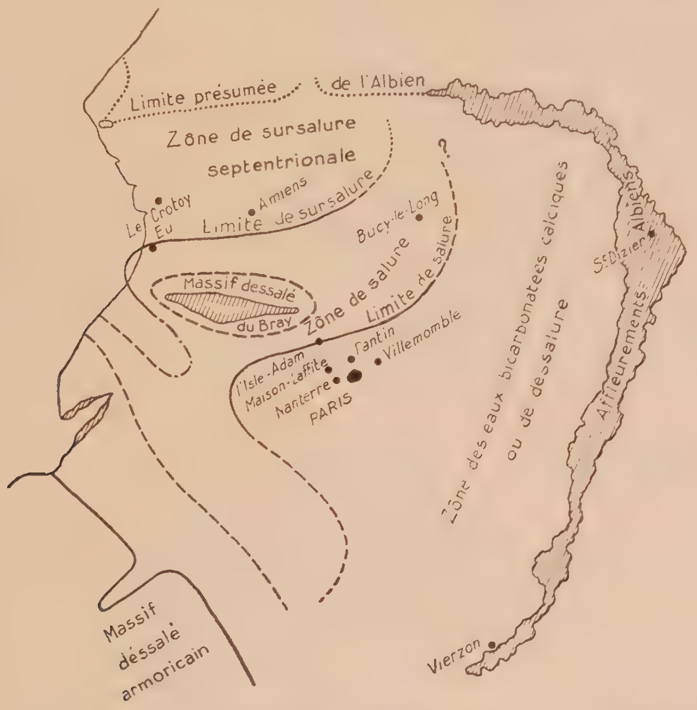

FIG. 7. — La nappe des Sables verts de l'Albien du Bassin de Paris d'après J. DELECOURT, *loc. cit.*

Enfin, si nous comparons la zone de salure et la zone de sursalure, à l'aide des forages de Bucy-le-Long et d'Amiens, nous constatons une forte augmentation de tous les éléments. Il y a concentration par dissolution, mais ici sans échange de base nouveau, car l'indice reste le même.

J'interprèterai les résultats de la manière suivante :

Les eaux de la zone d'alimentation (zone des affleurements de Delecourt) pénètrent dans la nappe captive et subissent, aussi bien dans la zone de dessalure que dans la zone des échanges de base, une très faible concentration et un faible échange de bases. C'est que les eaux y circulent activement. C'est somme toute la zone de circulation. Cette zone de circulation forme une pointe vers l'embouchure de la Seine. C'est donc dans cette direction que se fait le maximum de circulation d'eau, comme l'abaissement de la surface piézométrique l'indique d'ailleurs. Par contre, la circulation est vraisemblablement moindre vers la Somme.

Et si, latéralement et en aval de cette zone de circulation, c'est-à-dire dans la zone de salure de Delecourt, la concentration devient importante, les échanges de base considérables, c'est que l'eau n'y circule que lentement, laissant aux dits phénomènes le temps de se réaliser pleinement.

Plus en aval, dans la zone de sursalure de Delecourt, il n'y a plus échange mais simplement concentration. C'est que l'eau n'y circule pour ainsi dire pas. C'est une zone de stagnation. On arrive ainsi à une image semblable à celle que j'ai donnée en 1934 (1) des phénomènes chimiques et hydrodynamiques de la nappe des calcaires éocènes et de la nappe des calcaires sénoniens du synclinal de l'Oued el Kebir, en Tunisie.

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(1) H. Schoeller : « Les échanges de bases dans les eaux souterraines vadoses. Trois exemples en Tunisie » *Bull. Soc. Géol. Fr.* (5), IV, 504, 1934.

6) PROPOSAL TO CLASSIFY GENETICALLY THE THERMAL ITALIAN SPRINGS

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Italy is very rich of thermal springs : more than one thousand of them have been identified. They have given birth to a large number of studies; but these studies concern essentially the chemical composition of waters and their healthful effects. The private initiative, above all, for commercial reasons, had stimulated the researches in this field, adjourning by degrees the analyses of the most important thermomineral springs; and now we know the chemical and physico-chemical characteristics of many of them according to modern exigencies, thanks also to the « Committee for the study of mineral waters « founded by the National Council of Researches in his Committee for Chemistry. (1)

Researches are much less advanced for what concerns the waters origin, their temperature and mineralization, because of the scarce interest given to them by commercial and medical attendance concerns, the lack of a central organization, and the greater intrinsic difficulty of the problems to meet.

Lately very remarkable advances have been made about the genesis of two groups of springs such as : the sulphurous-saline waters of the Parmesan Appennines, connected with hydrocarbons formations (2), and the hypothermal Euganean ones, connected with volcanism. (3). The knowledge of the waters belonging to the first group (made deep above all by the petroleum and metanus researches) has allowed to know better the origin of a numerous category of springs that were not very known before. On the other hand, the sure presence of post-volcanic emanations in the Euganean thermal bath, together with the advanced knowledge about the phenomena which accompany the slow phases of the magma consolidation have validy supported the thesis long affirmed by B. Lotti (4) for the large system of thermal waters which spring in the metalliferous Tuscany.

We have deduced valuable elements out of these researches in order to classify genetically Italian thermal springs. For this purpose we have tried to examine critically data and hypothesis scattered in a very rich, but geologically rather obsolete, bibliography (5). Our work is rather far from the attempts made so far : (6) we present it with a certain trepidation because we know it is a rather imperfect sketch, but we wish it may be useful as a starting point for future researches.

I. — Sulphurous or saline, hypothermal or rarely mesothermal springs, connected with the genesis of petroleum.

1. — *Springs originated from gypsum formations.*

These are springs having generally a low temperature, coming out from gypsum (eventually also saline) formations which owe their temperature to the geothermic degree and their mineral contents to the washing out of the soils they cross in their way and to the reduction of the calcium sulphate by the work of the hydrocarbons.

They are often found in the Alpine as well as in the Appennine gypsum formationsl of every age. Their temperature is generally superior of only 1°-5° to the annuas average of their spring place. They are genetically connected with many otherwhich have not even this very weak thermality.

A series of sulphurous springs flow along the « permian » and « triassic » gypsiferous layers of the Venetian Alps : among the most known, we mention those of Bagni di Lusnizza, Arta, Villasantina, Calazo, Cimagogna di Auronzo, Valgrande in Comelico. Gypsum reduction is due to bituminous substances which are in the fetid limestones and rotten dolomits, including permian gypsum and in the marls including the « raibliani » gypsum (7). The water of such springs is entirely of a meteoric origin; usually the springings are situated along fault-lines.

The gypsiferous and sulphur formation of the upper miocene is found with a rich series of sulphurous waters which flow along the layers or through the pliocene or quaternary covering, along the Emilian Adriatic and Jonico side of the Appennines, some along the Tirrenico side, and a good number in central and meridional Sicily. The most known springs belonging to this group are the sulphur springs of Monferrato, of Broni and Stradella, of Imola, of Carignano and Montefelcino in the « Pesarese » region, of Aspio near Ancona, of S. Vincenzo la Costa and Strongoli in Calabria, of Enna, Caltanissetta, S. Cataldo, S. Caterina Villermosa, Racalbuto, Canicatti, etc. in Sicily.

They are often sulphurous-sulphured, owing to the easy gypsum solubility; the hydrocarbons apt to reduce the calcium sulphate are generally in the same gypsiferous and sulphur formation, but they may also come from sapropolic dregs placed below, often represented by the « flysch » of the » scaly clays «. In such a case it is possible there are passages from this category to the following one; then it is difficult to establish a definite separation. This happens particularly in the case, not very rare, in which the waters are not only sulphurous, but also saline (generally saline-brome-iodic) or with hydrocarburets: as it happens not only in the Northern Appennine, but also in the Abruzzo Chietino, in the Campano Appennine and sometimes in Sicilia. However, in this last region, very many sulphurous or saline-sulphurous springs are connected not only with the great extension and power of the gypsum-sulphur formations, but also with the bitumens which accompany the gypsum and saline marls of the upper miocene.

Among the few springs belonging to this group which have a remarkable temperature, due to the depth of the hydrothermal circuit, we must mention those of Palena in the province of Chieti (35°-48°), those of Caccuri (32°-33°) and Galatro (37°-39°) in Calabria, and those of Sclafani (33°-35°) and Termini Imerese (43°) in Sicily.

2. — *Springs derived from sapropelic formations.*

These springs have generally a low temperature and are genetically connected with others which have a normal temperature; however, there are some high temperatured ones belonging to this group. A common characteristic for all of them is to receive contribution (or better to derive) from geological formations apt to give origin to hydrocarbons; they are very often associated with their hydrocarburets. They are usually connected to dislocations or fractures and their waters are often of a mixed origin. Their strong or weak thermality derives from their geothermic degree and their ascensional power derives in the largest part from gas pressure.

This powerful system, consisting essentially of dark clays, which are laminated when not chaotic, eminently plastic, saline and more or less impregnate with liquid and gaseous hydrocarburets, must be deemed as a sediment of deep sea, originally sapropelic, and consequently the rock which gave origin to the hydrocarburets which now it contains.

The bonds of such formation with the sulphurous springs and the saline-bromeiodic ones of the Northen Appennine, even if they come out from younger grounds, are now demonstrated. Such springs occur in a large number along the Emilian zone of the Appennines, from the Vogherese (such as those of Salice, Rivanazzano, Retorbido, Casteggio), in the Parmense (as those of Salsomaggiore, Tabiano, Miano, Medesano, Fornovo, Lesignano) and in the « Modenese » and « Bolognese » regions (as those of Salvarola di Sassuolo, in many « salines » of Peglio, of Castel S. Pietro). And also towards the Appennine ridge, such as those of Bobbio and of the Porretta territory.

Researches have shown the connection with *diapiric* intrusions of scaly clay in a zone *tectonicaly* upset for what concerns « Parmensi « and « Vogheresi « springs. The fundamental constancy of the mineralization type is a proof that, even if the lixiviation of the crossed neogenetical ground has perhaps increased the quantity of the dissolved substances, the greatest part of them derives from the original rock. But probably their origin is much deeper; otherwise it would not be possible to explain the presence of a great quantity of boracic ion in the waters of Salsomaggiore (from 2 tot 2,5 gr/liter over a fixed residue of gr. 179) and in the other springs placed at the foot of the « Parmense » Apennine. Mr. Anelli has rightly put in connection the presence of the boracic acid with the *baryts*, the metallic sulphides and the other different minerals which give a proof of an hydrothermal activity of a magmatic origin in the Emilian scaly clay.

About this subject we must hint at the frequency of « datolite » in the argillous-ophiolithiferous formation of the Northern Apennine; but as in the mineral waters of which we are speaking, silex is found in very small proportions, the decomposition of that boron mineral should be excluded; and as a long durability of free boracic acid in the clays is rather impossible, we are obliged to think of the coming of *new waters* which contain it in solution and that so far continue to evolve (which should also explain the good quantity of ion Sr, contained in the waters of Salsomaggiore in the proportion of 0,2 gr/l). In such a case we should have a mixed origin for the waters as well as for their mineralization : *new waters*, *connate waters* and *external waters* which lixiviate the covering grounds.

Salsomaggiore waters appear on the surface at 16° : this low temperature may be explained by their long and difficult ascensional walk through the argillous scaly *lama* intruded in a pressed, faulted and curved anticlinal, and by the scarce permeability of the above layers. The conditions of the waters chemically similar of the Porretta are different; they spring out at 36°-38° from groups of miocene rocks fixed in an anomalous tectonic position in the scaly clay. The waters take their mineralization from these clays; their mineralization is similar to that one of Salsomaggiore, except for the concentration, which is much weaker (7 gr/l of fixed residue) and here also boracic ion is present, which makes us come to the same conclusions, the more so because it is accompanied with measurable percentage of barium and strontium. The easiness on hydraulic circulation in the fractured rock and its deep immersion in the argillous-scaly chaotic formation explain the great dilution (by a good contribution of meteoric waters), as well as their noteworthy thermality.

Probably a sub-stratum more or less deep of scaly clays is in the sector of « Romagnolo » Apennine included between the Sillaro and Marecchia Valleys, where numerous sulphurous and saline springs come out on the surface through a large and powerful fractured arenaceous-marly plate of the medium miocene, to which « messiniani » and pliocene grounds are superposed, towards the plain. So the warm sulphurous waters of Fiorenzuola, of Accerreta high Valley, of Bocconi in the high Montone Valley, the acid-saline waters of Bagno di Romagna (43°) in the high Savio; and also the hypothermal saline-iodic springs of Castel del Rio, Riolo dei Bagni, Brisighella, Rocca S. Casciano, Castrocaro, Bertinoro (La Fratta), Verucchio, all these become connected with the preceding ones and are disposed in a series parallel with them on the external margin of the Apennine. Their common origin is shown by the presence of the boracic acid in the waters of Castrocaro and Brisighella, though in a smaller proportion than in those found at Porretta and Salsomaggiore.

Spring waters of the same type make their way with difficulty, or have been found out by borings, corresponding to the folds sunk under the alluvions of the Po Valley. These are the waters of Miradolo, Castenedolo, Podenzano, Fontevivo (Fontanellato), Montepelato (10 Km. SE of Parma), Novi Modenese, etc., raising by the pressure of gassy hydrocarbous from aves contained in powerful medium Miocene marly sandy lensed masses, covered by newer grounds. We think that these are the masses which generated hydrocarburets and fossil saline-iodic waters, above all after that, by the big drilling of Montepelato, saline watery levels were found at a depth which varied from 850 to 2000 meters, with very strong pressures (from 200 to more than 300 atmospheres). But also these waters, which have a very similar composition to these of Salsomaggiore contain much boracic acid (gr/l 2,1 on 140 gr/l of fixed residue), besides the strontium; so that we must think also in this case, of deep contributions, and we think that the low temperature that the waters have when they spring out, is due to the loosing of heat during the very slow raising, which is so obstructed that the natural effluxion of the spring is made precarious and irregular, notwithstanding the intense gas pressure.

Many sulphurous-saline springs, in the most part hypothermal such as those of Penna S. Giovanni and of the Salino torrent in the province of Macereta, of Mozzano near Ascoli Piceno and of Rotella North of Monte dell'Ascensione, come out from a geological decidedly sapropelitic formation of the supper miocene (argillous marls with thin layers of sandstones), which extends from Macereta to Teramo; the abundant very sulphurous thermal waters of Acquasanta in the Piceno (35°-37°), discharge 60 l/s come out from the same grounds.

Also the sulphurous-saline springs of the Valle Latina and Agro Pontino, having a temperature from 2° to 6° superior to the annual average, are though to be dependent from the bituminiferous upper Miocene : we think that this bituminous impregnation is primary, though we agree with Ing. Beneo about the probable existence of another bituminous primary stratum under the big Mesozoic gypsum masses of the Central Appennine.

On the contrary the sulphurous (and also bituminous) mineralization of « vadose » waters under Marsico Nuovo and near Tramutola in the high Agri valley is connected with grounds of a « flysh « type.

II. — Sulphurous, ferruginous or ferruginous-arsenical springs connected with the decomposition of pyrites.

Also these are hypothermal springs. Notwithstanding that many authors, also modern ones, affirm that pyrites alteration process is a source of heating for springs having even a very high thermality, it is sure that waters which are surely mineralized in this way, have a temperature which is little superior than the local average, and sometimes they are completely cold.

We have instructive examples of them in Italy such as : the ferruginous-arsenical springs of Valsugana (Levico-Vetriolo, Roncegno, S. Orsola), of Staro and Monte Civillina of Vicentino, of Casino Boario in Valcamonica, of Tartavalle in Valsàssina, of Vanzone in Val d'Ossola, of Borgofranco d'Ivrea, and of Ceresole Reale at the foot of the Gran Paradiso : all these springs coming out from arsenious pyrite layers. And also the sulphate-ferruginous springs of Cavelonte in « Val di Fiemme » and of Serrai in « Val di Pine » ; and the ferruginous ones of Borgo in Valsugana, of « Oltre il Colle » in Val Serina and of Brosso near Ivrea, these last coming out from the mines drifts, as sulphurous waters of Boccheggiano and Massa Marittima do. The same thing happens for the acidulous-ferruginous-arsenical springs of Recoaro, though their contents are more complex, because they derive from various mineralization (but prevalently pyritic) that the « filoniane » basic rocks inducted in the mica-schists. Also the ferruginous arsenical springs of Gimigliano in Calabria take their mineralization (according to Cortese) from arsenious pyrites contained in « filladici » schists; however they have a rather high temperature (35°), because of the penetration of the hydraulic circuit in depth. A similar considération cán be made for the scarcely thermalized spring of Pré-St.-Didier in Val d'Aosta, which comes at 34^e, from metalliferous schists. Finally there is a very numerous category of sulphurous and ferruginous springs which usually have a low temperature and discharge. These springs come out from sediments of different nature and age, and they mostly flow through fractures or faults. Their temperature is due to the penetration of the circuit, that is, to geothermic degree; their mineralization is probably due to the reducing action of the ferrous sulphate formed by the oxidization of small quantities of iron sulphures spread in the sediments and melted by vadose waters.

A great number of them are mainly in the Alps and Venetian Pre-Alps, in Liguria and Toscana, and in the Umbro-Marchigiano Appennine. Problably the Acque Sante at Gerace Antonimina in Calabria belong also to this group. They are among the few waters which have a noteworthy thermality; they are saline-sulphate-ferruginous, gushing out in small springs from eocenic marly limestones at 34°-37°.

III. — Thermal springs connected with volcanism.

Springs gushing out from active or extinct volcanoes, or in their immediate surrounding zone and which draw water and mineral substances from magmatic vapours belong to this group. The composition of the waters and the geological conditions are as a guide to recognize such springs; there are cases of uncertain solution, above all in peri-volcanic zones, specially round about eruptive centres. They are very important in Italy; we mention the most important groups as follows :

1) — The Euganean thermal springs; they are saline-brome-iodic and lithic (with traces of boracic acid, strontium, arsenic), radioactive, springing out in pressure and at high temperatures up to 85°-92° along the « Colli Euganeai » from Abano to Battaglia and S. Elena (probably on the way of a fracture line) and also in the internal zone at Calzignano. The new waters and the gas, which accompany them, spread in the Artesian vadose layers met in the raising, warming and mineralizing them (G. Dal Piaz). The sulphurous springs of Barbarano between the Euganei and Berici have probably the same origin.

2) — The springs of the volcanic « Laziale » region, which are chiefly bicarbonate-sulphate – alkaline-earthy, radioactive, specially connected with the volcan Sabazio : for instance such as the thermal waters of Viterbo (38°-63°), Vucarello (48°), Stigliano (36°-56°).

3) — The thermal baths of Suio, which are sulphurous-sulphate – alkaline-earthy, coming out along the Garigliano, on the border of the volcanic cone of Roccamonfina (temperature from 16° to 39°; total discharge 3m^3/sec.; big proportion of vadose waters).

4) — The springs of Flegrean Fields, specially at Agnano (temperature from 20° to 105°) and Pozzuoli (from 40° to 67°), saline-bicarbonate-alkaline and saline-sulphate-alkaline; also arsenical at Bagnoli.

5) — The springs of Ischia Isle, saline-bicarbonate- (sulphate)- alkaline, gushing out all round the isle, at a temperature from 36° to 75°. A radioactivity corresponding to 1390 and 2850 Unity Mache, inferior in the world only to that observed in the waters in the mine of Jachimow has been measured in the springs less diluted by freatic water, at Lacco Ameno and Imbò thinks it is possible to reach greater values.

6) — The springs of the Vesuvian zone, sulphurous-saline-sulphate-alkaline. They are all with a low temperature and mixed with a great quantity of vadose waters : Torre Annunziata (26°), Valle di Pompei (14°), Castellammare di Stabia (15°-16°; discharge 190 l /sec.); these last waters with traces of lithium, strontium and barium.

7) — The springs of « Eolie » isles. We can mention : the bicarbonate-sulphate-clorurate-sodium springs of Bagnoscuro and S. Calogero (61°), besides various fumaroles in action, in the isle of Lipari; the ipothermal spring of Porto di Levante (88°) and the saline waters of Bagno di Vulcano, in the isle of Vulcano; a spring near Malfa (at 35°) and some submarine discontinuous fumaroles near Rinella, in Salina isle.

8) — The springs of the Aetnean zone. Besides the celebrated « Salinelle », the very acidulous-alkaline-ferruginous ipothermal waters of Paternò and the bicarbonate-sulphurous-saline-iodic sources of. S. Venera near Acireale, gushing out at 23° and independently of the hydraulic superficial variations are the only ones that we think directly dependent on volcanism.

9) — The springs of Pantelleria. Many of them are in the isle. Besides the fumaroles (« favare ») with aqueous vapour jets and other gas, there are the sulphurous and saline-sulphate-alkaline springs with a temperature up to 75° (S. Nicà); noteworthy the groups of Gadir (60°-70°) and Caldarelle surrounding Bagno dell' Acqua.

10) — The springs of Sardegna. The greatest part of numerous thermo-mineral springs of Sardegna receive their aliment from post-eruptive phenomena, which are a consequence of the intense volcanic neogenic and quaternary activity of the isle. Such, for instance, the bicarbonate-alkaline waters of Sàrdara (5 springs at 50°-68°, radioactive, with a discharge of 12 l /sec, Fordongianus, Abbarghente, Bonorva; the clorurate springs of Villasor (62°, 6 l /sec); the sulphurous-clorurato-alkaline springs of Benedutti (34°-46°) and of Casteldoria (70°-75°); the bicarbonate sulphate-alkaline-earthy springs of Siligo, Florinas, Thiesi, Ploaghe : these last ones coming out also at a very low temperature (15°-20°) but, for their origin and composition, they are not separable from others.

IV. — Springs connected with post-eruptive emanations

The springs belonging to this group are generally and also totally vadose waters. They have their origin as ascensional springs by hydrostatic pressure (usually along fault-surfaces) and also by gas. Their mineralization derives more or less from the soils they cross but the gas is of a magmatic origin. This gas may also contribute to rise their thermality. This is very different according to various cases and can be often explained with the depth reached by the hydrothermal circuit. Gas and vapours derive from the consolidation of more or less deep magmas either by volcanic warmth or by intrusive bulks and they can get rather far from their origin place. It is natural anyhow that both the areas near volcanoes and the regions in which we have lately had a development of hydrothermal phenomena (filonians or through impregnation, etc.), are more or less rich of thermal springs.

Toscana offers a very instructive example about such kind of phenemena, that B. Lotti has already well shown. There the greatest part of thermal waters are localized in great number in the area of the ancient « metalliferous chain », between the Appennine and the sea, from Lucca to Civitavecchia. The most important groups of springs follow each other from NW to SE in the following orders :

a) — Montecatini and Monsummano baths (32°-35°, saline-iodic-bicarbonate-sulphate-alkaline-earthy, radioactive springs, with Li, P, NH_4 and a great percentage of nitrogen in the gas).

b) — Ponte Pisano springs : their waters are sulphate-alkaline-earthy at S. Giuliano Baths (33°-41°); bicarbonate-alkaline-earthy at Uliveto (23°-34°, of a constant discharge, rich of CO_2, with Li and He, greatly radioactive) and having a lower temperature (20°-25°), also at S. Giuliano and besides at Agnano, Vicascio, Calci.

c) — Casciana Baths (36°, sulphate-alkaline-earthy, lithiferous waters, rich of nitrogen).

d) — Boracic « soffioni » of Val di Cecina and Val di Cornia (temperature from 100 to 12° in the natural « soffioni », which rises to 190° in the drillings; with CO_2 mixed to H_2S, H_2, N_2, CH_4, etc.), and the thermal sulphurous waters of the Galleraie (or of Radicondoli, 49°) and of Pomarance (42°-44°) connected to them.

e) — The springs of the Valley of the Marse (Petriolo, Macereto, Montisi, etc., sulphurous waters, 27°-45°) and of Boccheggiano (36°-45°, with boracic ion).

f) — Campiglia Marittima springs, sulphate-alkaline-earthy waters (31°-43°), complexive discharge 150 l /sec.

g) — Gavorrano Baths (34°, sulphate-alkaline-earthy waters, discharge 80 l /sec.)

h) — The springs of Caldarella, of Poggetti and of Roselle between Montopescali and Grosseto (35°-40°), sulphate-alkaline-earthy waters.

i) — The group of M. Amiata, with the springs of Saturnia (sulphurous waters, 35°-37°, discharge 400 l /sec), S. Filippo springs (25°-53°, sulphurous-bicarbonate-sulphate-alkaline-earthy waters, with Sr, near a cinnabar mine), the springs of Vignoni (49°-52°, sulphurous-bicarbonate-sulphate-alkaline-earthy waters).

l) — The group of M. di Catona, with the springs of Rapolano and Motalceto (29°-40°, very sulphurous-carbonic waters), Chianciano (29°-39°, bicarbonate-sulphate-alkaline-earthy waters) and of S. Casciano dei Bagni (43 springs at 31°-42°, sulphate-alkaline-earthy waters, discharge 62 l /sec).

m) Bagno dell' Osa near Talamone (32°) and Bagni di Canino (39°) north of Civitavecchia.

Such groups of springs (8) are in relation with faulting and thrust faulting in an old mountain area (Catena Metallifera Toscana).

The springs are situated along the fault lines some of them are visible and well determined, some others only supposable according to the disposition of the blocks or of the groups of fractured and differently raised blocks. We notice in the same area : 1) the quaternary pouring rocks (trachytes of Val di Cecina, of Roccastrada, of Monte Amiata, etc.); 2) a great number of filonian fillings, impregnations, alterations and big travertine plates, which demonstrate an intense quaternary hydrothermal activity, continued till very lately; 3) post-eruptive phenomena still in activity, such as in particular many emanations of CO_2 and H_2S, that are in connection with the boracic « soffioni « demonstration of a magma cooling, not very deep. All this shows how the modern thermal springs of metalliferous Tuscany are the attenuated continuation of lately post-eruptive phenomena and in large part they owe their amineralization and according to cases, some contribution of heat and new waters to them. We must notice that quaternary eruptive and posteruptive Tuscany phenomena have been preceded by very remarkable intrusions and effusions of similar magma (granites and trachytes of Campiglia Marittima, granites of Gavorrano and Elba), referred to the upper miocene and connected with the intense hydrothermal activity to which we owe the metalliferous « filonians » and metasomatic layers of the region.

Another group of carbonic-sulphurous springs, on whose origin the various opinion agree, is found in the surroundings of the Lazial volcan. The most important are the Albule Waters of Tivoli (23°), discharge 4 m³/sec, with their contents of boracic acid, a great quantity of CO_2 and also of H_2S and their strong encrusting power.

But we cannot deny a similar origin to the similar and not less important carbonic-sulphurous springs of the Valley of The Velino, among which those of the plain of S. Vittorino and Paterno excel; though 65 Km. far from the nearest eruptive centre. The waters of the celebrated Mefita di Ansanto (35°), with the near thermal baths of Villamaina (36°) in the territory of S. Angelo dei Lombardi in the province of Avellino, are also far from volcanic zones and rich of CO_2 and H_2S whose origin is surely endogenous.

Another example is offered by the hydrothermal springs of Telese in the Valley of Volturno, more than 40 Km. far from the Vesuvio and from Roccamonfina. These waters are sulphurous-saline-bicarbonate-calcic, radioactive, very rich of CO_2 and H_2S, with lithium, strontium, iodine, brome and boracic acid, having a low temperrature (20°) due to their large mixing with a big quantity of meteoric waters absorbed by calcareous massive. The springs of Contursi (22°-40°), in the basin of the Sele, have the same characteristics; they are very sulphurous and sulphurous-saline and boric and lithic.

These facts induce us to more general considerations.

The narrow relation between magma consolidation and thermo-mineral springs, which we could ascertain in Tuscany for a whole of favourable circumstances, and which extends logically to the zones of recent volcanism, is not limitated only to these territories in Italy. Quaternary post-orogenetical movements in which the deformations and the mutual displacements and rolling of the various blocks have resolved in a general (though very various) raising, through phases of light subsidence and also accentuated subsidence. The intensity and extension of the eruptive phenomena which have accompanied these movements in the Tirrenic and circum-Tirrenic area, shows an intimate bond between such movements and magma activity; and with reason Salamon assumed of magmatic raising, that is of pushes due to magma ascensional strenght, in order to explain the very recent raising of the Calabrese Appennine, and so afterwards R. Signorini did for metalliferous Tuscany. Consequently we must think that magmatic activity has a much greater importance and extension in our Country than generally we admit and the preceding emanations of the magma consolidation are still in action in many places of Italian region, from the Alps to Sicily.

The most important examples are offered, in the Alpine zone,by Acqui, Vinadio and Valdieri. The sulphurous-saline-sulphate springs of Vinadio and Valdieri are scarcely mineralized, like many others of the granitic and gneissic massives; but their high temperature, which raises till 68° in both the groups, and the lithium and phosphoric acid they contain, make us think, as G. Roverero says, that there are communications between the deep part of the crust of the earth and the external part.

The group of sulphurous and sulphate-saline springs of Acqui, where the temperature raises up to 71°,5, is connected with the gneissic massive of M. Laione (coming out on the surface at 5 Km. SSE of Acqui) and to the serpentines which accompany it, and which are also altered by the hydrothermal action with abundant production of magnesium sulphate. Besides these facts, remembered by G. Rovereto, we must add that the waters contain lithium, strontium, boracic ion, and a very high percentage of nitrogen in the desolved gas. Besides these springs we must mention the many other sulphurous waters connected with the serpentines of « Green stones » Ligurian zone. The most known of them are the saline-sulphate-sodium waters of Acquasanta di Voltri (26°).

Always going on in the Alpine zone, we think that also the thermal baths of-Bormio belong to this group. These waters are bicarbonate-sulphate-alkaline earthy, very radioactive; they come out at 37°-42° by the contact through the fault of powerful calcareous block with the schistous crystalline formation; and also the alkaline waters coming out from the gneissic rocks of the Thermal baths of Masino in Valtellina (38°) and of the Thermal baths of Craveggia in the Ossola(28°-31°). It is probable that also the baths of the Brennero belong to this group. Their waters (22°) are scarcely bicarbonate-sulphate and very much radioactive, and the Thermal baths of Comano in the Trentino (27°), oligomineral, radioactive, rich of N_2 and CO_2; they contain barium and come out from a very fractured field, in a zone of large fault and tectonic rolling.

In the Lombard Pre-Alps, we must mention the very sulphurous (and also saline-iodic, radioactive) springs of Trescore Balneare and Zandobbio (13°-17°) and the scarcely mineralized, radioactive waters, containing lithium, of Bracca and S. Pellegrino (27°). All these coming out on the surface along contact superficies between principal dolomitic and « retici » schists in zones tectonically upset. We find the very sulphurous, and radioactive springs of Monfalcone (39°) and S. Stefano d'Istria (36°), much more set towards the East; they are connected with the great fractures of the calcareous cretaceous formations and their deep base.

We have the waters of the thermal baths of Lucca (35^e-54^e) which are sulphate-alkaline-earthy and radioactive, and of The Garfagnana (Equi, Torrito di Castelnuovo, Pieve di Fosciana, 30°-40°) saline-sulphate, having common characteristics with the thermal waters of the « Metalliferous Chain », in the « Apuane » Alps. Around this, there are the big bicarbonate-calcic, very incrusting springs of Colle Val d'Elsa (22°, discharge 500 l /sec) and Sarteano (24°,120 l /sec), the Acqua Borra near Siena (35°-37°), saline-sulphate-alkaline) waters; the very known bicarbonate-sulphate-alkaline-earthy springs of Chianciano (23°-29°) and, Southwards, the thermal baths of Civitavecchia (56°), having a similar composition, which perhaps are already in connection with the « Laziale « volcanism.

The springs of Monteluinno, near Gubbio, in the Umbro-Marchigiano Appennine are very rich of H_2S; the waters of Sangemini and Villasanfaustino (Massa Martana) are very carbonic : near the latter spring we must notice the little thermal spring of Casigliano, associated with a powerful emanation of H_2S. We must admit similar emanations fot the very sulphurous waters of Triponzo (Spoleto) coming out at 31° from a well delineated fault; for the big very sulphurous and carbonic springs of Acquasanta near Ascoli (35°-37°, containing traces of As, Ba, Sr, Rb, Cs, Li, P and a remarkable quantity of nitrogen), similar to those not very far from Castel Trosino (17), 380 1 /sec), and for those very sulphurous ones of lower Aterno coming out below Raiano with 500 1 /sec. of discharge.

The very sulphurous waters of Santa Cesarea (27°-30°) flow in Terra d'Otranto. They are quite isolated on the fractured margin of the powerful crestaceous tableland : waters certainly of vadose origin but with a considerable gaseous contents of CO_2, H_2S and N_2, that notwithstanding the various tests already done, is inexplicable if we don't think of deep emanation.

Some springs which come out from crystalline and schistouscrystalline blocks of Calabria are even more instructive about this subject. Above all, the Luigiane Thermal Baths at Guardia Piemontese (Cosenza) sulphurous-saline-brome-iodic springs with a temperature reaching the 45° that Talarico has demonstrated are of a mixed origin and connected with a mineralized quartz layer. The similar springs of Spezzano Albanese (Cosenza) at 24°-28°; the sulphurous ones of Cerisano and of Cassano in the Cosentino (26°), of Bivongi and Sambiase in the Catanzarese (21°-40°); the sulphate springs of Condofuri in the province of Reggio, belong to this group.

Finally a rich series of springs that we think are mineralized, and also warmed by endogenous emanations, is found in Sicily. The sulphurous-saline thermal baths of Ali Marina (from 18° to 46°) connected with the schistous crystalline Peloritan massive; the very sulphurous ones of Castroreale (31°); the sulphurous and sulphurous-saline springs of Termini Imerese (43°, radioactive), Sclafani (35°), Castellammare del Golfo (44°-49°, discharge 200 1 /sec), Alcamo (74°), Segeste (30°-48°) and Calatafimi (49°).

V. — Scarcely mineralized springs, not connected with mining layers and depending exclusively on hydrothermal circuit.

Tha waters which are neither sulphurous nor carbonicate, neither saline nor ferruginous, flow along an hydrothermal circuit and they owe their temperature to it and their dissolvent action to their mineral contents. Springs on this kind are rather rare in Italy and consequently not much known. We mention, with due reserves (as examples) the abundant springs of S. Nazario (28°, 200 1 /sec) on the banks of Lesina lake, north of Gargano; some springs of Solofrana torrent (High Sarno), coming out from gypsum massives at 21°-25°; some springs coming out from crystalline schists in the « Catanzarese » Region (Uria torrent) at 22°-25°; various springs not mineralized of Western Sicily, such as Acqua Calda near Trabia(S. Nicolò l'Arena, 26°), Cefalù Diana thermal baths (west of the former one, 38°); Acqua Calda of Montevago (Agrigento), coming out from limestones at 31° with a constant discharge of 113 1 /sec; Bulgherano and Acqua Calda di Grammichele, from basaltic grounds of the Northern Iblei (22°-25°), and the big springs of Beverotto and Sigona Grande, coming out from basaltic tuffs at 23° with respective discharges of 54 and cf 41 1 /sec.

Conclusion.

Italian thermal springs may be grouped, for what concerns their genesis, as follows :

I. — Sulphurous or saline hypothermal or rarely mesothermal springs connected with the genesis of petroleum;
   1) derived from gypsum formations
   2) derived from sapropelitic formations.

II. — Sulphurous, ferruginous or ferruginous-arsenical springs connected with the decomposition of pyrites.

III. — Springs connected with volcanism.

IV. — Springs connected with post-eruptive emanations.

V. — Scarcely mineralized springs not connected with mining layers and depending exclusively on hydrothermal circuit.

This last group offers the least number of examples and on which there is the greatest uncertainly.

The above stated scheme is a summary attempt, bent, above all, to state the present knowledge of the matter and pointing out the working field for future researches.

NOTE

(1) Mineral and thermal springs were partly an object of study by Ing. Zoppi and Perrone in the « Memorie illustrative della Carta Idrografica d'Italia » (Ministero di Agricoltura, Industria e Commercio, Roma 1888-1920); temperature and discharge measurements were accomplished for many of them by « Servizio Idrografico » del Ministero dei Lavori Pubblici, as is shown by Servizio Idrografico Issues, N° 18, III, Italian Springs, Roma, under compilation (the fascicles 1-7 already published).

(2) See above all M. Anelli, Geological Hints about Traversetolo and Lesignano Thermal baths surrounding, *Boll. Soc. geol. It.*, 34, 1915, p. 79;

Id., Mineral Water of Montepelato, La Min. It., 38, 1925, p. 201;

Id., Mineral waters in the hills between Stirone and Taro, Giorn. It. Sc. idromin. and clim., I, 1930, pags. 145 and 182.

(3) See De Marchi, Origin of Thermal waters of Montegrotto, Rend. Acc. Lincei, (6) 5, 1° sem., 1927, p. 841 and la Memoria of G. Dal Piaz presented to this Congress.

(4) B. Lotti, Geological mining description of Massa Marittima surroundings, *Mem. descr. Carta geol. D'It.*, 8, 1893; Id., Die geol. Verhåltnisse der Thermalquellen in Toskanischen Erzgebirge, Z. f. prakt; Geol., 1893 p. 372; Id., About Hydrological conditions of Thermal springs of Vignoni near S. Quirico d'Orcia, *Boll. Com geol. It.*, 26, 1895, p. 219, and Z.f. prakt. Geol., 1895, p. 372; Id., Tuscany Geology, Mem. descr. Carta geol. d'It., 13, 1910; Id. Thermal spring of Bagno di Gavorrano, *Boll. Uff. Geol.*, 56, 1931, mem. N° 4.

(5) See M. Gortani, General Bibliography of underground waters in Italy, Publ. N° 28 del *Serv. Idrogr.* It., fasc. II, 1938.

(6) These reduce to the note of E. Perrone, About the origin of thermal mineralized waters (in *Mem. ill. Carta Idrog. d'It.*, N° 31, 1904, p. 365) and of A. Trentin, Geographic distribution of mineral springs in Italy *(Riv. di Idrol. e Climatol.*, 37, 1926, p. 326).

(7) Since very often also ferruginous waters come out from alpine gypsum we must think that in such cases the reducing action of ferrous sulphate derived from pyrites oxidation or marcasite spread out in the sediments, has perhaps contributed to the liberation of H_2S.

(8) The number of such groups might be increased if we enumerate some less important ones such (as for instance, acidulous waters of Livorno and of S. Giovanni d'Asso, the sulphurous ones of Pereta and of S. Gimignano, etc.) or if we include some conspicuous but scarcely or not at all mineralized springs, such as those, of Massa Marittima (22° - 25°, about 1500 l/sec.).

7) LES EAUX MEDICINALES DE LA HONGRIE

par

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Les eaux médicinales de la Hongrie au point de vue hydrogéologique se répartissent :

1°) en eaux d'origine postvolcanique,
2°) en eaux minéralisées par la décomposition ou
3°) par le lessivage des minéraux que comportent les couches argileuses et limoneuses,
4°) en eaux médicinales de caractère mixte jaillissant des couches profondes.

Conditions géologiques

Dans l'évolution géologique du pays un rôle important fut joué par les mouvements orogénétiques de la chaîne Varisque ainsi que des Alpes et des Carpates, le développement du bassin de la Grande et Petite Plaines hongroises y compris.

La chaîne Varisque est constituée 1°) en Transdanubie : *a*) des massifs granitiques de *Velence* (15) (*), *b*) et du promontoire SE de la montagne *Mecsek* (17) tandis que 2°) sur les bords nord de la Grande-Plaine hongroise elle est représentée par la montagne *Bükk* (50). Sur les bords SO de cette dernière, ce sont le gabbro et le diabase, et, sur les bords E et NE, la fillite, les schistes argileux et le calcaire de l'époque carbonifère qui émergent.

(*) Voir les chiffres sur la fig. 1.

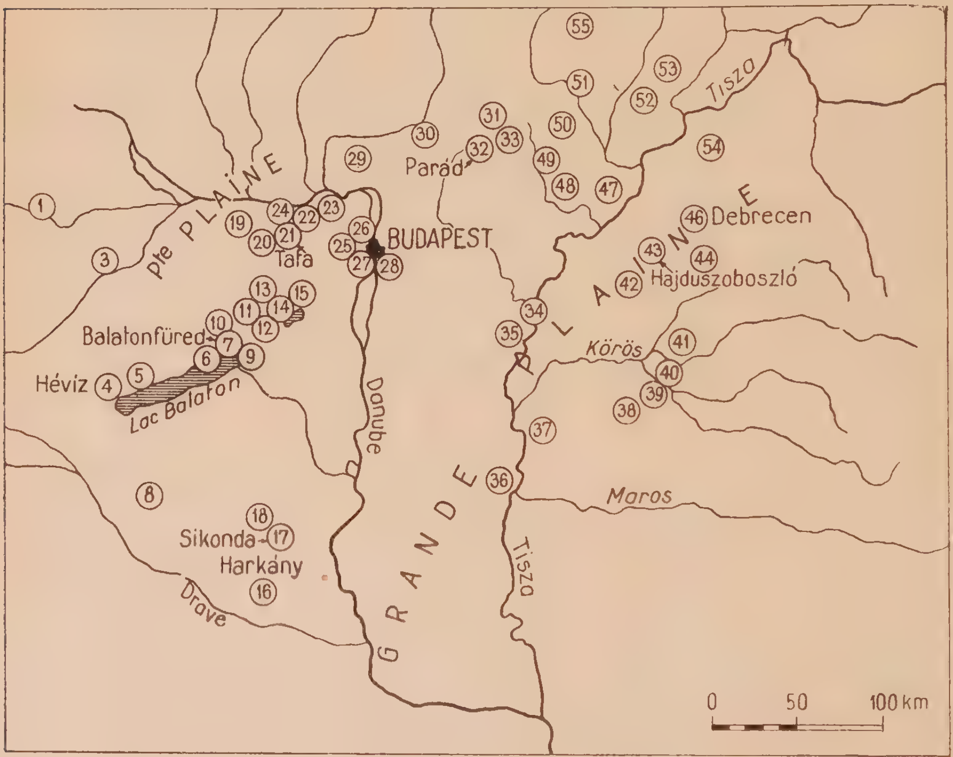

FIG. 1. Lieux d'émergence des eaux minérales et thermales de la Hongrie.

Le granit de la montagne de *Velence* accuse une vive activité hydrothermale s'étant déroulée à l'époque carbonifère et mésozoïque. L'effet en est la présence des filons de barite, fluorite et stilbite (zéolite), ainsi que de calcite au contact de granit et des dykes d'andésite. Dans le granit de la montagne de *Mecsek* on retrouve également des filons de quartzite, des venules de kaolin ainsi que des minéraux différents : fluorite, hematite et calcite comme produits postvolcaniques. A l'heure actuelle c'est uniquement sur les bords SE de la montagne de *Velence* aux environs de *Székesfehérvàr*, que les eaux thermales émergent du granit dans les sondages profonds. Le massif de granit et diorit de *Velence* s'étendit sur tous les environs de *Budapest*, puis démembré par suite des mouvements orogénétiques il s'effondrait dans les niveaux plus profonds. C'est à cette façon que le dit massif joue un rôle dans le chimisme des eaux thermominérales de *Budapest* et ses environs.

De même que les roches enfoncées de la montagne de *Velence* font sentir leur présence et influence sur la composition chimique des eaux thermominérales émergeant des régions avoisinantes, les massifs granitiques du *Mecsek*, dans ses parties SE influent directement sur la composition des eaux thermales des bains de *Sikonda* (18) voire celles de *Harkàny* (16).

Dans les roches de gabbro et de diabase de la montagne de *Bükk* (50) des minerais sulfurés (calchopyrite, pyrite), epidote et quartzite accusent l'activité ancienne hydrothermale. Les eaux émergeant aux environs de la ville de *Eger* (49) et de *Mezőkövesd* (48) ont également dissous les dites roches. En dehors des frontières du pays, c'est surtout dans la région de la montagne de *Szepes-Gőmőr* (55) que l'on retrouve les anciens lieux d'émergence des eaux thermales indiquées par la présence des mines comme p. ex. à *Rozsnyó, Csucsom, Szomolnok, Dobsina.*

La chaîne Varisque s'étendant autrefois sur la superficie totale du pays, s'est démembrée par suite du plissement des chaînes alpines-carpathiques et sa majeure partie s'effondrait profondément.

Le développement des Alpes et des Carpates a été accompagné d'une vive activité volcanique. Les traces de cette activité sont, en premier lieu, le tuf d'andésite et la dacite-andésite ayant fait une irruption à travers de celui-ci ou s'étant répandu sur sa surface au miocène, tandis que l'époque pliocène est marquée par des filons et des nappes de basalte et exceptionnellement par le tuf de basalte. Les éruptions d'andésite et de dacite sont accompagnées d'émergences de sources thermominérales. Une série de filons de divers minerais témoigne de la dite activité dans les chaînes bordant la *Grande-Plaine.* Les éruptions de basalte ont été suivies de la zéolithisation et de la naissance de l'aragonite. Les exhalaisons (solfatares, mofettes), de cette récente région volcanique minéralisant les eaux des bains de *Balatonfüred* (7) et de *Paràd* (32) etc.

Ce sont non seulement les sédiments de l'époque carbonifère et mésozoïque, mais encore les roches éruptives du tertiaire qui ont été disloquées en sens vertical aussi bien qu'horizontal, tandis que les roches compactes sédimentaires (dolomites, calcaires, les couches de marne et de grès) se sont morcelées et plissées.

Nous signalons comme exemple par rapport à la dislocation verticale caractéristique : la dolomite de trias de Budapest s'étant enfoncée au-dessous de la superficie jusqu'à 917 m à une distance de 1,5 km du mont St. Gellért. A *Hajduszoboszló* (43) le sondage atteint les roches mésozoïques à une profondeur de 2017 m. On retrouve le fillit du massif mésozoïque à une profondeur de 138 m au bord du *lac Balaton.* La Grande et la Petite-Plaines entourées par les bassins de moindre importance sont constituées de diverses couches de roches sédimentaires et les eaux phréatiques y cheminant contiennent, à maints endroits, des éléments chimiques très appréciés.

I. — Eaux minérales d'origine postvolcanique

Les exhalaisons postvolcaniques liées en Transdanubie aux éruptions de basalte et, sur les bords septentrionaux de la Grande-Plaine, aux éruptions de l'andésite et de basalte transforment, à maints endroits, la couche supérieure de la nappe phréatique en eaux minérales.

1°) *Eaux minéralisées par des mofettes.*

Une partie de nos eaux minérales contenant de l'anhydride carbonique est d'émergence naturelle, mais on y en rencontre également des sources ouvertes par sondage.

a) Les sources naturelles à teneur de l'anhydride carbonique se trouvent sur la rive NO du lac *Balaton.* L'anhydride carbonique se mélange avec la couche supérieure de la nappe phréatique à une profondeur de 6-7 m au-dessous du niveua du sol. Au-dessous de la surface, on rencontre un terrain d'origine de marais de 3 à 4 m d'épaisseur avec des couches de tourbes intercalées. Puis, au-dessous de celui-ci, on relève l'argile sablonneuse du pliocène inférieur. Il y a des endroits où le grès rouge de l'époque permienne émerge. Dans quelques puits c'est par les fissures du grès que s'effectue l'exhalaison de l'anhydride carbonique (70 à 850 cm^3 par litre). Le débit des sources est de 25 à 30 litres/min. en moyenne. (La température en est en moyenne 11 à 12° centigrades, celle de la région est de 10,6°.)

On utilise les eaux d'anhydride carbonique pour des bains thérapeutiques depuis la moitié du 18e siècle (*Balaton-füred*, 1743) (7). Il y a lieu encore de noter *Kékkut* (5) et *Moha* (13).

b) Loin des régions constituées de roches volcaniques, les sondages profonds de recherche pour hydrocarbures ont approché les centres éruptifs d'où les témoins de l'activité postvolcanique, tels que les vapeurs, gaz, se dégagent. C'est l'anhydride carbonique qui jaillit abondemment en maints endroits.

Aux environs du lac *Balaton*, les sondages ont percé, au-dessous des couches du pliocène inférieur, une série très variée de couches du trias. Elle peut être évaluée, en Transdanubie à 1300 à 1950 m d'épaisseur, dont la moitié (800 m) consiste en dolomite (4 séries de couches), puis, ce sont les couches de marne qui y succèdent en ordre de la fréquence et de l'épaisseur (600 m, 4 séries de couches). L'épaisseur de la couche de calcaire est plus faible (350 m environ). Les failles aussi bien que les plissements déterminent l'endroit des exhalaisons de CO_2. C'est au croisement des failles de même que le long des anticlinaux que les exhalaisons de l'anhydride carbonique sont les plus fortes. Les émergences déjà mentionnées : *Balatonfüred*, *Kékkut* et *Moha* correspondent aux dits endroits.

L'exhalaison jaillissant à *Mihályi* (3) est un cas tout à fait spécial. C'est là, que la Société A, Hungaro-Américaine d'huiles minérales a obtenu, en 1937, une quantité de 150,000 m^3 de l'anhydride carbonique par jour d'une profondeur de 1510 à 1525 m et de 1550 à 1557 m. Le sondage a été exécuté au-dessus d'un maximum géophysique. Plus tard, il apparut, — sur la base des résultats du sondage —, que l'on est tombé sur un important laccolite de basalte dont l'anhydride carbonique d'origine postvolcanique se frayait passage.

Les exhalaisons de l'anhydride carbonique sont connues non seulement aux environs du lac Balaton, mais encore sur les versants nord de la montagne de *Mátra*. Aux alentours de *Parád* (32) on a observé des exhalaisons d'anhydride carbonique. Au sud des bains qui s'y trouvent dans la vallée *Ilona*, on a pénétré jusqu'à une profondeur de 21 m soit par creusement, soit par sondage, et on a atteint, au-dessous des couches de gravier de l'alluvion, le biotite amphibole andésite dont les fissures accusent un abondant débit de l'anhydride carbonique. L'andésite est fort décomposé démontrant les traces d'une kaolinisation, silification et minéralisation. On a introduit dans le puits de la mofette les eaux de la nappe phréatique.

2°) *Eaux minérales solfatariennes.*

A Parád (32), dont nous avons fait mention ci-dessus, des eaux solfatariennes jaillissent également. Les éruptions d'andésite ont été suivies d'une vive activité postvolcanique. L'andésite tantôt a kaolinisé, tantôt des quartzites se sont formés. Parmi ces roches altérées, il y a lieu de noter, l'enargit, la pyrite, la calchopyrite, galène et la blende, etc., comme minéraux méritant une attention au point de vue de leur extraction. L'ion K se trouve en très faible quantité dans les eaux des bains solfatariens de *Parád* (0,001 g/litre), par contre, les ions Na et Ca sont considérables (Ca = 0,1348 g/litre), fait dû aux plagioclases de l'andésite. Les eaux contiennent par litre 0,0499 g de SO_3 et 1,435 g de CO_2, par conséquent elles sont sulfuriques, acidulées, ferrugineuses et alcalines. Aux environs de Parád, en rencontre, à plusieurs endroits, des sources analogues à celle de Parád, comportant, cependant, moins de substances dissoutes.

Le débit des sources acidulées et sulfuriques est de 70,000 litres par jour.

II. — Eaux minéralisées par la décomposition des minéraux des couches argileuses et vaseuses

(Eaux sulfatées, eaux sodiques)

Sur le périmètre S et SO de *Budapest* ainsi qu'à *Igmánd* (19) et *Jászkarajenő* (35) dans les couches argileuses, se trouvant près de la superficie se détachent des sulfates. L'acide sulfurique de la décomposition de la pyrite, dispersée dans l'argile et la vase (provenant des combinaisons sulfuriques des organismes défunts) attaque les minéraux facilement désagrégeables de Ca, Mg et Na, de sorte que la nappe aqüifère devient bicarbonatée sodique. (Voir le rapport présenté par M. *A. Vendl* relatif aux eaux minérales sulfatées de Budapest.)

A Budapest il y a quatre groupes importants de sources d'eau minérale sulfatée :

1°) groupe *Hunyadi János* (actuellement 10 puits, leur nombre était de 179 avant la guerre).
2°) groupe *Ferenc József*, (34 puits),
3°) groupe *Apenta*, (29 puits),
4°) *Erzsébet Sósfürdő*, (17 puits alimentant surtout des bains thérapeutiques).

Aux environs d'*Igmánd* (19) on trouve également des sources d'eaux minérales sulfatées. Leur occurence est analogue à celle des sources de Budapest. Il y a 23 puits enfoncés dans l'argile de l'oligocène. On exploite à présent 10 puits. La moyenne de la profondeur des puits est de 5-6 m, la température des eaux est de 14°, leur moyenne de débit est 4900 litres par jour et par puits avec une quantité des sels dissous de 17,5 kg.

Il en est de même à *Jászkarajenő* (35), village situé dans la région NO de la Grande-Plaine où les sols alcalins sont fréquents. Il y existe 17 puits, la profondeur moyenne desquels est de 2 à 4 m. L'alluvion consiste en sol alcalin au-dessous duquel se trouve le diluvion, puis viennent des couches d'argiles vaseuses comportant l'*epsomit*, qui provient de la décomposition des pyrites. Les eaux de *Jászkarajenő* contiennent $Na_2SO_4 = 38{,}127$ g, $MgSO_4$ ... 16,995 g par rapport à 1000 g. Le débit des puits est de 4800 litres par jour.

Les puits d'eau minérale purgative de Budapest et des autres régions mentionnées fournissent au total un débit de 12,600 litres par jour.

A *Jászkarajenő*, de même qu'à maints endroits de la *Grande-Plaine*, les eaux alcalines des étangs des anciens champs d'inondation sont utilisées, depuis des siècles, surtout contre les diverses maladies opiniâtres de peau. Les étangs alcalins sont alimentés par les eaux des couches supérieures de la nappe phréatique. Les eaux des précipitations atmosphériques stagnantes dans les dépressions du terrain plat attaquent les minéraux facilement décomposables du sable et de la vase, à savoir les débris fins de la calcite, dolomite, biotite et du feldspath. C'est de ces minéraux que les sels alcalins proviennent par suite des réactions suivantes :

$$Na_2SO_4 + CaCO_3 = CaSO_4 + Na_2CO_3$$

$$2\,NaCl + CaCO_3 \updownarrow Na_2CO_3 + CaCl_2$$

$$2\,NaHCO_3 = Na_2CO_3 + CO_2 + H_2O$$

Des lacs contenant Na_2CO_3, $NaH/CO_3/$, Na_2SO_4 et NaCl se trouvent à *Kakasszék* (38), *Nyiregyháza-Sóstó* (54), *Konyár* (44).

III. — Eaux minéralisées par suite du lessivage des sels des couches argileuses et vaseuses

A maints endroits de la Grande-Plaine les eaux de la nappe souterraine sont minéralisées par la dissolution des sels des terrains d'origine lagunaire.

Les couches argileuses-vaseuses de l'oligocène et de miocène à *Pesterzsébet* (faubourg de Budapest) ainsi qu'à *Sóshartyán* (30) et à *Tiszagyulaháza* (47) accusent une riche teneur en sel gemme lessivé par les eaux souterraines. Les eaux de 24,5° centigrades du puits de *Pesterzsébet*, provenant d'une couche sablonneuse de gravier de quartz situé dans la profondeur de 100-103 m, contiennent 11,89 g de NaCl et sont utilisées pour des buts thérapeutiques. Le débit journalier est de 228,000 litres.

A *Sóshartyán* (30) l'eau d'un puits de 134 m de profondeur pénétrant dans les couches d'oligocène contient 13,07 g de NaCl ainsi que 22,60 mg d'iode et 92,31 mg de brome (d'origine organique).

Aux environs de *Tiszagyulaháza* (47), on a trouvé dans les eaux d'un puits enfoncé dans l'alluvion 11 g de NaCl par litre et les eaux des puits des environs à 20-30 km sont de nature analogue. On suppose que les eaux alcalines émergent le long des failles dans les régions N et NE de la Grande-Plaine. Le débit des puits salins est de 880,000 litres par jour.

IV. — Eaux thermales émergeant des couches profondes de caractère mixte

On connaît sur le territoire du pays 129 sources hyperthermales d'une température supérieure à 30° C, et 137 sources thermales tièdes (20-27° centigrades).

Ces dernières sont des eaux de Carso mélangées avec des eaux émergeant des couches profondes. Il y a des endroits où les sources tièdes émergées dans le pleistocène et le miocène ont complétement tari et ce sont uniquement leurs dépôts qui témoignent l'endroit de leur activité d'autrefois. Le meilleur exemple en est *Tihany* (6) démontrant à peu près 70 cônes de geyser bien séparés les uns des autres, geysers perçant les couches argilo-sablonneuses de pliocène inférieur. Les dépôts calcaires des anciennes sources accusent la présence de calcedoine, de Fe_2O_3, voire même de l'or. Des dépôts calcaires sont connus également au NO du lac *Balaton* à *Várpalota* (10), *Inota* (12), *Szomod* (20). Les travertins des carrières de *Süttő* (24), *Dunaalmás* (22) et *Békásmegyer* (27) le long du Danube donnent d'excellentes pierres de construction bien que l'on ne trouve plus les sources tièdes dont ils ressortent.

Les sources tièdes émergent à présent en partie dans leurs endroits primitifs naturels, le long des failles d'une manière si abondante qu'elles alimentent des lacs ou des ruisseaux, en partie elles sont ouvertes par des puits ou par des sondages profonds. Leurs eaux de 20 à 27° centigrades sont utilisées pour des buts thérapeutiques ou industriels. A *Tata* (21) l'eau tiède (*lac Cseke*) émerge à un niveau absolu de 140 m, au bord du Danube, au niveau de 130 m, tandis que, plus loin, à Budapest, la hauteur absolue du niveau d'émergence est de 102 m.

a) *Emergences naturelles d'eau tiède.*

Les émergences naturelles d'eau tiède sont marqués par la présence d'un lac ou ruisseau. Bien que le débit des sources soit considérable, il accuse des fluctuations sensibles. Il y a deux facteurs exerçant une influence sur le débit des sources minérales tièdes se trouvant sur le territoire de *Budapest*, à savoir : la pression atmosphérique et la hauteur du niveau d'eau du Danube. Quant au débit des sources émergeant à *Tata*, on a pu démontrer l'influence de la pression atmosphérique et de l'attraction de la lune.

Voici quelques données relatives aux débits (valeurs moyennes) :

Source :	Température centigrades	Débit litres/min
Budapest		
Rómaifürdö (Bains romains)	20-21	10,000
Source Arpád d'Obuda	20	2,000
Lukácsfürdö (Bains St. Luc)		
Malomtó	24	2,100
Source Romain	24	2,100
Császárfürdő		
Source Turque	21-29	5,500
Sziklaforrás	25	990
Tata (21) (débit total	20	288,000
Eger (49)	28	14,000
Pét (11)	22,5	48,000
	Total :	373,190 litres-minute

soit 54,000 m³ par jour. En adoptant un résidu solide de 0,7 g par litre, la quantité journalière des sels dissous est de 38 tonnes en somme globale.

Le débit de ces sources est également influencé par les facteurs extérieurs susmentionnés ce qui peut être le mieux illustré par les résultats des nombreuses mesures relatives aux sources *Arpád* d'Obuda et de *Romaifürdő* de Budapest.

Le graphique (*fig.* 2) relatif à la source *Arpád* fait ressortir que l'influence de la hauteur d'eau du Danube sur le débit de la source est presque insignifiante, étant donné qu'elle se trouve à 1,6 km du Danube. Par contre, on peut constater l'influence directe des variations de la pression atmosphérique. Le débit s'accroît ou s'affaiblit suivant que la pression atmosphérique diminue ou monte, fait vérifié par mes propres mesures et celles de M. *Tarics.*

La fluctuation du débit du groupe de sources de *Rómaifürdő* (*fig.* 3) démontre que l'influence du Danube exercée sur le débit n'est pas sensible en dépit de sa proximité (800 m). Les hauts niveaux du Danube, cependant, font sentir leur effet

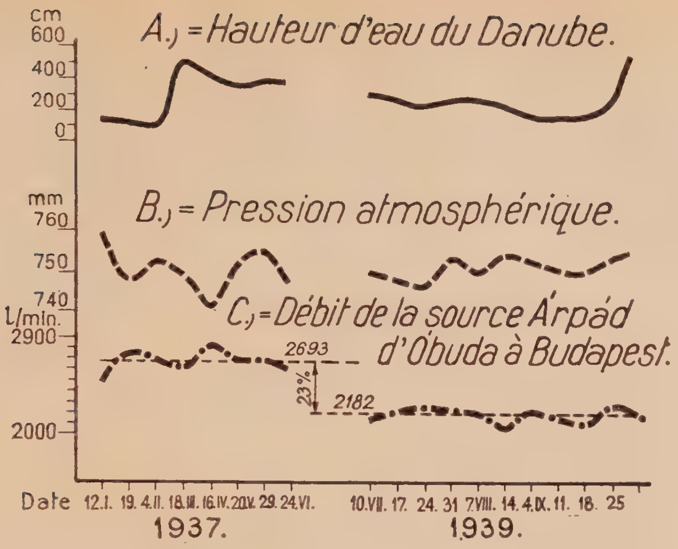

FIG. 2. Variation du débit de la source Arpád d'Obuda à Budapest.

après deux ou trois jours (par exemple le 14 juillet 1947). Aux mois de septembre et octobre de 1947 la pression atmosphérique et l'attraction de la lune s'expriment nettement dans le débit. C'est surtout la baisse du débit des sources s'étant présentée entre les 23 et 28 du mois de septembre qui accuse l'effet de la haute pression atmosphérique (761,9 mm).

Si le débit des sources thermominérales varie, leur température n'accuse que des

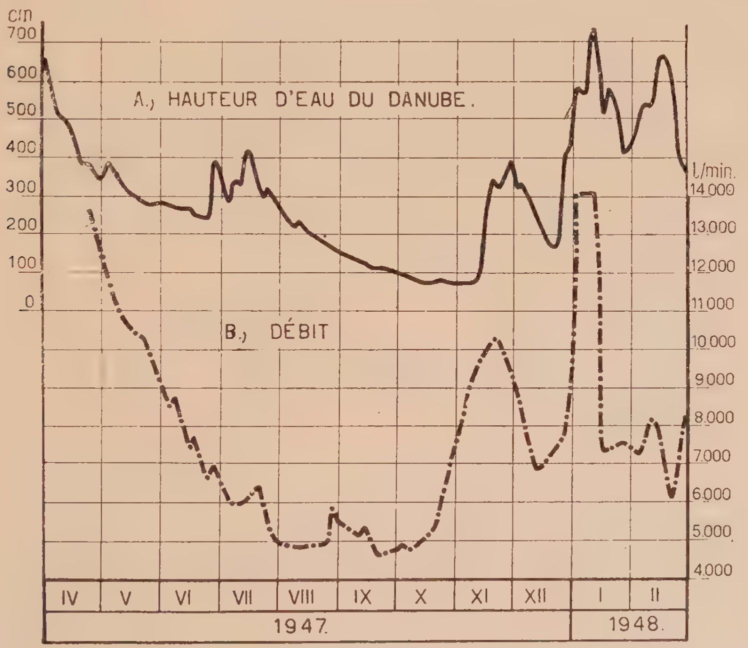

FIG. 3. Fluctuation du débit des sources de Rómaifürdő de Budapest en fonction de la hauteur d'eau du Danube.

différences du dixième de degré centigrade au cours d'une année. Les observations relatives à une période de 12 ans démontrent par contre, dans le cas de la source Arpâd, une baisse de 2,1° C, changement dû à l'épuisement forcé de la source par une usine industrielle qui provoqua l'infiltration croissante des eaux froides de Carso.

Les sources tièdes émergent des diverses roches. Par exemple les sources de *Romaifürdő* (Bains romains) de Budapest sourdent des couches sablonneuses d'alluvions du Danube, les sources de *Tata* (21) sortent de couches de sable du pliocène inférieur, tandis que les sources de *Eger* (49) jaillissent des couches argileuses-sablonneuses de l'oligocène. Les eaux tièdes ne proviennent pas effectivement, dans ces cas, des dites couches mais des niveaux plus profonds. Une partie des eaux se caractérise comme l'eau des calcaires et des dolomites triasiques, tandis que l'autre partie provient des précipitations atmosphériques, enfin une partie est formée par la déshydratation des roches se trouvant dans les niveaux très profonds. Suivant les analyses les eaux des sources tièdes naturelles sont terreuses, bicarbonatées, sulfatées, et radioactives. (Voir *Relevé* I.)

b) *Sources minérales tièdes ouvertes artificiellement.*

On a réussi à ouvrir, aux failles thermales et synclinaux, des sources d'eau tiède à l'aide des sondages profonds ou des puits creusés.

Les puits artésiens d'eau tiède les plus importants sont :

Budapest
Rómaifürdő, Ivóforrás (puits creusé)
  Csillaghegy 3 puits.
  Profondeurs : N° I : 109 m, N° II : 137 m, N° III : 150 m.
Pünkösdfürdő (26) 533 m.
Dunaalmás (22) (2 puits creusés).
Esztergom (23) 323 m.
Eger (49) 3 puits : 43,4 m, 248 m, 228 m.
Görömbölytapolca (51) 8 puits creusés.
Mezőberény (40) 420 m.
Szentes (37) 330 m.
Békés (39) 733 m.
Szolnok (39) 967 m.

Abstraction faite de quelques exceptions, les forages ont, partout, atteint les assises du calcaire mésozoïque, de sorte que ce sont les eaux émergeant de ces roches qui alimentent en premier lieu les puits susmentionnés. Il y a lieu de signaler cependant, les eaux froides de superficie ainsi que les eaux thermales en provenance de la déshydratation des roches des niveaux profonds qui participent également à alimenter ces sources. La proportion dans laquelle les eaux d'origine différente se mélangent peut être déterminée d'après leur température.

La hauteur absolue du niveau actuel d'émergence des sources est à *Romaifürdő* (à Budapest) de 103 m, à *Mezőberény* (40) 100,6 m, à *Szentes* (37) 101,3 m. Il y a lieu de noter qu'à Romaifürdő on retrouve le travertin indiquant l'endroit de l'émergence d'autrefois à 1,5 km des sources actuelles et à un niveau supérieur de 47 m au niveau d'émergence actuel.

Le débit des puits énumérés est considérable :

Budapest, Csillaghegy	17,000	litres-minute
Pünkösdfürdő (26)	1,800	» »
Dunaalmás (22)	1,200	» »
Esztergom (23) Puits St. Etienne	1,400	» »
Görömbölytapolca (51)	20,800	» »
Mezöberény (40)	125	» »
Szentes (37)	780	» »
Total, en chiffre rond,	40,000	litres-minute

En additionnant à cela le débit des sources naturelles, on obtient comme *débit total* 400,000 *litres-minute soit* 600,000 *m*^{3} *par jour.*

Bien que le débit des sources énumérées soit variable, leur température est pratiquement constante, ce qui prouve que les eaux froides de la nappe phréatique ne se mélangent pas avec les eaux des sources. Par suite de la pression des eaux tièdes émergeantes, les eaux de la nappe phréatique ne peuvent pas pénétrer dans les griffons même dans le cas où leur niveau est très haut. Par contre, si le niveau des eaux phréatiques est bas, les eaux tièdes alimentent la couche supérieure aquifère.

Relevé I. — Minéralisation des eaux médicinales tièdes de la Hongrie.

	Budapest										
	Bains Rómaifürdó	Sources Arpâd d'Obuda	Bains St. Luc Malomtô	Bains Csâszárfürdö Source Turque	Csillaghegy++	Pünkösdfürdó ++ (26) +	Tata (21) +	Dunaalmás ++ (22) +	Esztergom++ (23) +	Görömbölytapolca (51) +	Eger++ (49) +
		1000 g d'eau contiennent									
K	0.0014	0.0027	0.00491	0.0156	0.0016	0.0013	0.00202	0.0068	0.0034	0.0018	0.00142
Na	0.0142	0.0169	0.02146	0.1249	0.0155	0.0167	0.00843	0.0209	0.0097	0.0017	0.00925
Ca	0.1040	0.1075	0.10025	0.1241	0.0997	0.1101	0.08732	0.0876	0.0964	0.1143	0.10196
Li	—	—	0.00010	0.0008		trace	trace	trace	—	0.0001	trace
Ba	—	—	—	0.0006		—	—	—	—	—	—
Mg	0.0421	0.0402	0.03837	0.0426	0.0404	0.0421	0.04766	0.0477	0.0423	0.0098	0.00723
Fe	trace	0.0022	—	0.0012		0.0002	0.00008	—	0.0042	—	
Mn	trace	—	0.00020	0.0008	—	—	—	—	0.0005	0.0005	
Cl	0.0075	0.0025	0.04599	0.0418	0.0093	0.0115	0.00770	0.0191	0.0213	0.0105	0.00386
H_2CO_3	0.4737	0.4910	0.34426	0.7392	0.4661	0.4875	0.23115	0.4763	0.4373	0.3894	0.32910
SO_4	0.0641	0.1448	0.11034	0.1155	0.0533	0.0669	0.03920	0.0498	0.0474	0.0234	0.02946
SiO_3	0.0161	0.0068	—	0.0221	0.0172	0.0151	0.01033	0.0238	0.0423	0.0592	0.01359
Σ	0.7231	0.8146	0.93828	1.2306	0.7031	0.7526	0.43391	0.7321	0.7318	0.6052	0.49587
CO_2 libre	0.0634	0.0294	0.29240	0.1927	0.121	0.0185	—	63.3 cm³	0.1991	0.2334	0.16220
Ra em.	0.5016 mgRa /sec	0.054 mg Ra /sec	1.602 mg Ra /sec	0.51 mg Ra /sec	0.2641 mg Ra /sec	0.39×10^{-6} mill. curie	0.0464 mg Ra /sec	0.56×10^{-9} mill. curie	—		2.2×10^{-6} mill. curie

+ Les chiffres correspondent à la *Fig.* 1.
++ Puits ouverts par sondage.

La minéralisation des eaux des puits ressort du *Relevé I.*

La température des eaux, — mesurée toujours au même endroit, — est constante. Par contre, on relève des températures différentes dans le bassin de captage de 4,4 m de longueur de la source Turque des *bains Császár*, si on les mesure simultanément à divers endroits. En effet, on relève à l'extrémité sud et nord du dit bassin respectivemnet 21,3° et 29,4° C. Ce phénomène s'explique par l'influence de la source naturelle St. Etienne (60° C.) émergeant à 43 m au NE de la source en question. Ses eaux prennent fuite vers le dit bassin.

Les sources énumérées ont atteint, dans la plupart des cas le calcaire triasique. (*Pünkősdfürdő* (26), *Békásmegyer* (27), *Esztergom* (23).)

A *Eger* (49) on a retrouvé les eaux tièdes dans les assises plus récentes, soit en tufs de riolite du miocène, soit dans les couches d'oligocène. L'auteur est d'avis que ces eaux proviennent également des calcaires triasiques, ce qui est vérifié par leur composition chimique. On est à même de constater que les sources tièdes naturelles et artificielles sont, dans la majeure partie des cas des eaux de Carso auxquelles les eaux de la nappe phréatique sont mêlées.

Il est sans doute que les sources tièdes se trouvent en activité depuis le pliocène du fait que leurs dépôts (tuf de calcaire) se trouvent 40 à 50 m (à Budapest même à 120 m) au-dessus du niveau de leurs émergences actuelles.

Les sédiments démontrent, en outre, que le débit des sources, à l'époque pliocène, était supérieur au débit actuel. Le maximum du débit fut atteint dans le pleistocène diminuant depuis ce temps.

2°) *Les eaux hyperthermales du pays.*
sont en partie des sources naturelles, en partie ouvertes artificiellement.

a) *Les eaux hyperthermales naturelles* émergent dans les griffons se trouvant le long des failles (*Budapest, Héviz* (4), *Harkány* (16).) Les dépôts dans les griffons permettent de constater, sans exception, la présence de l'aragonite, pyrite, marcasite, barytine et du gypse (à *Budapest* : Bains St. *Gellért*, la source Turque des *bains Rudas*, la source St. Jean des *bains Császár*). Il y a des sources dont le sédiment est vaseux. En les examinant de plus près, il appert qu'il s'agit de granules cristallisés du dolomite du trias. Il y a des cas où les eaux thermales émergent par des alluvions (sable et vase exempts des substances colloïdales).

En ce qui concerne la température des eaux, c'est la source St. Antoin des *bains Császár* à Budapest qui est la plus chaude (63,4° C). La température des sources thermales varie d'environ 3 à 4° C s'adaptant avec un retard de 3 à 5 semaines, aux variations saisonnières de la température de l'air. La variation de la température dépend en outre de la quantité des vases sédimentées dans le bassin de captage. Plus la quantité des sédiments est grande, plus la température de l'eau diminue, phénomène très bien observé dans la source St. Antoin des bains de Császár.

Le débit des sources varie suivant le niveau des eaux de superficie. A Budapest, le débit des sources dépend de la hauteur d'eau du Danube. (*Voir Relevé II*). Plus celle-ci est grande, plus le débit des sources est abondant, sans que la température des eaux soit diminuée. Si le niveau du Danube est bas, une partie de l'eau thermale se perd dans le lit du fleuve. Inversément, si le niveau du Danube est élevé, le niveau de la nappe phréatique monte également, et les eaux froides (donc plus dures) entravent la fuite des eaux thermales.

Sur le territoire du pays, 24 sources thermales naturelles émergent dont il revient à Budapest 22. La cause de l'ascension des eaux est la pression hydrostatique et l'expansion des gaz dissous (surtout méthane) et l'acide carbonique. Leur niveau d'émergence correspond à peu près au niveau du Dabube (hauteur absolue de l'étiage 96,6 m), comme il ressort des données suivantes :

	Niveau d'émergence	Débit litres-minute
Bains Rudas, Source Mathias	104,34 m	108
Bains Király	104,00 »	360
Bains St. Luc, Iszaptó	102,00 »	1200
Bains Császár, Source St. Antoin	99,91 »	557
» » , Source Marie	102,98 »	500
» » , Sources St. Jean	102,73 »	166

On peut considérer ces sources comme puits artésiens naturels. Leurs eaux jaillissent, ou bien elles doivent être élevées par pompage et amenées à l'endroit de leur utilisation.

Relevé II. — Données d'observation relatives à l'influence du niveau du Danube sur le débit des sources thermales de Budapest

Source	Espacement du Danube m	Date du jaugeage	Débit de la source litres par minute	Date	Hauteur en cm du maximum le plus proche de la hauteur du niveau du Danube
Budapest Bains St-Gellért					
Source N° 1	50	28.III. 1947	2600	25.III. 1947	759
»	»	28.XI. 1947	1561	28.XI. 1947	398
»	»	14.VII. 1947	1457	14.VII. 1947	425
»	»	18.11. 1948	2265	15.11. 1948	683
Budapest *Bains Rudas*					
Source Mathias	60	22.II. 1943	101,4	19.II. 1943	284
»	»	22.VI. 1943	117,0	20.VI. 1943	589
»	»	7.X. 1943	84	3.X. 1943	269
Budapest *Bains St-Luc*					
Malomtó	280	17.IV. 1943	2634	6.IV. 1943	417
»	»	7.V. 1943	2448	28.IV. 1943	372
»	»	30.VI. 1943	2610	21.VI. 1943	589

b) *Sources hyperthermales ouvertes artificiellement.*

Les sources hyperthermales naturelles émergent le long des failles ou des synclinaux. Quant aux puits forés, on ne peut relever dans tous les cas ces formations caractéristiques. Les forages n'ayant pas fourni les résultats voulus démontrent cependant que les conditions d'émergence sont analogues à celles des sources naturelles. En effet, il y a des endroits où les sondages exécutés entre deux sources jaillissantes ont abouti à ouvrir des sources, dont les eaux ne jaillissent pas ou bien elles n'ont fourni aucune eau. Par exemple sur l'île Ste Marguerite à Budapest, deux forages effectués entre la source nord (43° C) et la source sud (76°C) n'ont abouti à aucun résultat.

Les gaz jouent dans l'ascension des eaux des sondages profonds un rôle très important. (*Hajduszoboszló* (43), *Bükkszék* (50), *Karcag* (42).)

Les eaux émergeantes sont de caractère mixte. Sur le territoire de Budapest, les eaux hyperthermales en provenance de la déshydratation des roches très profondes remontent le long des failles thermales et se mélangent, d'une part, avec des eaux tièdes de Carso (14 à 15°) provenant des assises de la montagne calcaire et dolomite située sur la rive droite du Danube, et, d'autre part, avec les eaux froides (6 à 10° C) souterraines, s'infiltrant de la surface. Entre les trois sortes d'eaux, les eaux de Carso sont les plus abondantes.

Parmi les sources ouvertes artificiellement on relève les sources les plus chaudes du pays, comme il ressort du *Relevé III.*

Le *Relevé IV* indique le débit de quelques sources hyperthermales.

Parmi les substances constitutives ce sont les chorides, bromides, iodides et hydrocarbonates qui méritent une attention spéciale et qui proviennent du lessivage des sédiments salins de la mer néogène (ayant occupé la Grande-Plaine dans l'éocène, oligocène, miocène et pleistocène). Les puits fournissent en outre des gaz appréciés (émanation du radium, méthane, etc.).

La minéralisation des sources en question ressort du *Relevé V.*

Une propriété très intéressante de ces sources est leur radioactivité. Ce sont, surtout, les sources des bains de Rudas de Budapest qui s'en distinguent considérant que la faille thermale n'est pas très profonde et que la roche la plus ancienne, la dolomite du trias supérieur, est proche de la superficie. La radioactivité des dites sources est analogue à celle des bains de Gastein. Les sources de 40 à 45° C sont plus radioactives que les sources tièdes ou hyperthermales. (Ces dernières jaillissent des sondages.)

Relevé III. — Température des sources hyperthermales de la Hongrie.

Budapest :	
Bains Széchenyi, Source St-Etienne	79,5° C
Ile Ste-Marguerite, Puits artésien N° 11	69,0 »
Bains Csészâr, Source St-Antoin	63,0 »
Bains St-Luc, Source St-Luc	62,0 »
Hajduszoboszló (43)	74,0 »
Karcag (42)	70,0 »
Mezókövesd (48)	70,0 »
Lâbod (8)	70,0 »
Debrecen (46)	65,0 »
Harkâny (16)	62,0 »
Szolnok (Bains de la ville) (34)	54,0 »
Szeged (36)	50,0 »
Tiszaörs	48,0 »
Nagyatád (8–b)	45,0 »
Békés (39)	43,0 »
Szeghalom (41)	43,0 »
Bükkszék (31)	39,4 »
Héviz (4)	38,0 »
Sikonda (18)	35,6 »
Eger (49)	32,0 »

Relevé IV. — Débit, en litres/minute, des sources hyperthermales les plus importantes de la Hongrie

Sources hyperthermales de Budapest	19,263
Bükkszék, puits N° 27 (31)	1,070
Bükkszék, puits N° 52 (31)	2,280
Réviz (4)	6,000
Sikonda (18)	1,500
Harkány (16)	1,500
Lábod (8)	700
Csokonya-Visonta	350
Szolnok (34)	500
Mezôkövesd (48)	5,000
Hajduszoboszló (43)	2,000
Debrecen (46)	2,600
Karcag (42)	2,000
Békés (39)	250
Szeged (36)	570
Nagyatád (8b)	280
Kaba	240
Cegléd i	2,000
Mezôberény (40)	125
Szarvas	250
Szeghalom (41)	230
Tótkomlós	220
Tiszaörs	300

Total : 49,228 litres/minute

soit 70,000 m³ par jour.

Relevé V. — Minéralisation des sources hyperthermales les plus importantes de la Hongrie,

	Budapest												
	Bains Széchenyi	Puits No 11 de l'île S. Marguerite	Bains St. Luc	Bains Császár	Bains Rudas	Bains St Gellért	Harkány (16)+	Hadjuszoboszló (43)+	Debrecen (46)+	Karcag (42)+	Bükkszék (31)+	Héviz (4+)	Sikonda (18)+
	1000 grammes d'eau contiennent :												
K	0.0160	0.0429	—	0.0187	0.0172	0.0274	0.0202	0.0213	0.0470	0.005	0.0646	0.0047	0.0515
Na	0.1702	0.0916	—	0.1394	0.1732	0.1699	0.1763	1.8240	1.7645	2.050	8.2436	0.0004	0.1132
Li	0.0002	0.0002	0.0002	0.0006	0.0005	0.0016	0.0003	0.00035	0.00025	—	trace	—	0.0019
Ca	0.1597	0.1480	0.1560	0.1598	0.1964	0.2008	0.0567	0.0016	0.0166	—	0.0872	0.0764	0.0857
Sr	0.0036	0.0043	—	0.0020	0.0005	0.0021	0.0046	—	—	—	0.0025	—	—
Ba	trace	—	—	—	—	—	—	—	—	—	0.0012	—	—
Mg	0.0355	0.0365	0.0448	0.0388	0.0591	0.0606	0.0127	0.0019	0.0049	trace	0.0323	0.0350	0.0358
Fe	—	—	—	0.0011	0.0011	0.0013	—	trace	0.0010	—	trace	0.0020	—
Mn	0.0004	0.0071	—	—	—	—	0.0021	trace	—	—	trace	—	—
Al	—	—	—	—	—	—	—	0.0042	0.0051	—	—	0.0004	—
Cl	0.1911	0.1233	0.0589	0.1613	0.1805	0.1940	0.1091	1.8189	1.8331	3.008	3.1040	0.0424	0.0112
HCO_3	0.5562	0.4961	0.5452	0.5843	0.6217	0.6189	0.5505	1.5231	1.7080	0.0121	16.9964	0.2606	0.6212
SO_4	0.2023	0.1481	0.0279	0.1880	0.3579	0.3866	—	0.0020	0.0109	trace	0.0075	—	0.1773
SiO_3	0.0660	0.0385	—	0.0547	0.0395	0.0536	0.0586	0.0780	0.0637	0.0899	0.0351	0.0609	0.0364
CO_2	0.5980	0.4087	—	0.2526	0.3009	0.2772	0.0126	—	—		0.7235	0.2250	—
Ra em.	0.09 mgRa /sec	$0.15.10^{-6}$ mill. curie	43.32 mgRaEm / sec	0.55×10^{-6} mill. curie	108.5 Eman / /lit.	37.35 Eman / lit.	$0.36.10^{-9}$ mill. curie	$1.71.10^{-7}$ mill. curie	—			56.06 Eman / lit.	$0.034.10^{-6}$ mill. curie

+ Les chiffres correspondent à la *Fig.* 1.

Les bains thermaux de *Budapest* et de *Hajduszoboszló* (43) sont très fréquentés. Les bains de *Debrecen* (46) sont non moins renommés.

Bien que les sources hyperthermales naturelles accusent, une variation de température de 3° C, les sources artificielles n'en démontrent que de 0,3 à 0,4° C.

De même que le débit des sources naturelles de Budapest est exposé à des fluctuations, le débit des sondages varie aussi suivant la pression atmosphérique. Cette corrélation peut être bien observée dans le cas des sources Juventus et Attila des *Bains Rudas* (*fig.* 4).

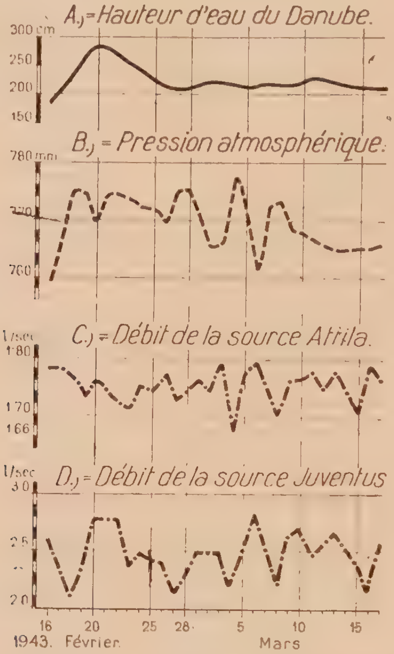

FIG. 4. Variation du débit des sources Attila et Juventus des Bains Rudas de Budapest.

Résumé

Sur le territoire du pays, les eaux phréatiques et les couches profondes sont, à maints endroits, minéralisées.

I. — La minéralisation de la nappe phréatique peut s'effectuer comme suit :

a) Les eaux météoriques infiltrées, stagnantes, comportant toujours une certaine quantité de CO_2, dissolvent des sels alcalins (*eaux sodiques*). La température de ces eaux peut atteindre même 30° C.

b) L'eau, dans sa circuite souterraine, lessive les sels gemmes du sol (*eaux salines*).

c) L'eau, par suite de la décomposition des pyrites des couches vaseuses-argileuses, devient sulfatée, (*eaux sulfatées, eaux sulfatées sodiques*). La température de ces eaux est de 14 à 16° C bien qu'elles soient alimentées par des eaux de superficie. Le fait qu'elles accusent une température supérieure à la température moyenne de l'air doit être attribué à la chaleur se dégageant au cours de la décomposition de la pyrite.

Avant la guerre, la Hongrie exportait une quantité annuelle de 5 à 6 millions de litres d'eaux minérales purgatives « *Hunyadi János* », « *Ferenc József* », « *Mira* » « *Apenta* ».

d) Les exhalaisons postvolcaniques (en premier lieu : anhydride carbonique et hydrogène sulfuré) minéralisant la nappe phréatique fournissent également des eaux médicinales très appréciées. Leur température est de 6 à 7° C, inférieure à celle des sources alimentées par des eaux de superficie, du fait de l'effet refroidissant des gaz qui se dégagent de l'eau. Ces sources ont été les plus actives dans l'époque pliocène, leur activité, va, depuis ce temps, toujours en décroissant. La plus célèbre en est *la source acidulée de Balatonfüred* (7) ayant un effet thérapeutique très avantageux contre les maladies de cœur. et des nerfs. Les bains de *Balf* (1), ville de climat subalpin, sont également très favorisés. Les bains de *Parâd* (32) au milieu de vastes forêts de hêtres sont exceptionnellement efficaces.

II. — *a*) Les eaux tièdes de Carso émergent des assises situées à une profondeur de 700 à 800 m. Les eaux de ces sources contiennent, en état dissous des sels de Ca, K et Na et des gaz (Ra, méthane, CO_2, H_2S) de sorte qu'en raison de cela, elles sont considérées comme eaux minérales. Les dépôts calcaires témoignent que les sources ont commencé leur activité dans l'époque pliocène. Elles étaient les plus abondantes dans l'époque pléistocène et, depuis ce temps, leur débit va en diminuant.

A *Tata* (21), les sources médicinales tièdes alimentent 21 lacs limpides entourés de bois magnifiques. Des eaux analogues sourdent à *Esztergom* (23), ville située sur le Danube, puis aux environs de *Budapest* : à Csillaghegy, Rómaifürdö, *Pünkősdfürdő* (26), enfin, à *Eger* (49), ville située dans un milieu pittoresque aux bords de la Grande-Plaine.

b) Les sources hyperthermales, les mieux appréciées de la Hongrie, jaillissent de profondeurs considérables (1300 à 1500 m). C'est surtout en raison de leur température très élevée (73 à 79° C) qu'elles méritent en dehors des substances y dissoutes (Si, P, Cu, As, Pb, F, B, J), une attention tout à fait spéciale.

Ces sources émegent depuis l'époque triasique. Leur activité était la plus forte aux époques pliocène et miocène. Leur débit est encore, à l'heure actuelle, exceptionnellement abondant.

Les bains hydrothermaux hongrois sont très recherchés par les malades du pays et de l'étranger. La situation de *Budapest*, où on relève 123 sources thermominérales, est unique même en relation mondiale. Ses sources thermominérales alimentent copieusement 14 grands établissements de bains et il y en a dont la température est de 70° C fournissant 11,000 litres-minute.

Notons encore sur la Grande-Plaine : les bains de *Hajduszoboszló* (43) et de *Debrecen* (46). Enfin en Transdanubie : les sources émergeant à *Héviz* (4), à *Sikonda* (18) et à *Harkány* (16), bains très efficaces surtout contre les maladies rhumatismales.

8)

SOURCES OF HEAT, WATER SUPPLY, AND MINERAL CONTENT OF STEAMBOAT SPRINGS, NEVADA (1)

by

DONALD E. WHITE and W. W. BRANNOCK

Introduction

Interest in the geology and origin of thermal springs of the United States was first apparent only a little more than 100 years ago. Most of the literature on the subject concerns individual springs or groups of springs but a few general reports have been published. The most recent and most inclusive of these is by Stearns, Stearns, and Waring (2). The reader is referred to that paper for a summary of the general problems and for an extensive bibliography on the thermal springs of the United States. Major problems concern the sources of heat, water supply, and mineral content. In general, investigators have shown marked disagreement on criteria to be used in the solution of these problems as well as upon their conclusions.

Geologists of the world have been interested in steamboat Springs, Nevada, since 1888, when Becker (3) stated that mercury, antimony, gold, silver, copper, and other metals were being deposited from the thermal waters. The main purpose of the present investigations of the United States Geological Survey is to determine the origin of these thermal springs and any bearing they may have on processes of ore deposition. Preliminary results of these and other phases of the investigations are being published elsewhere (4).

The springs are 19 kilometers south of Reno, Nevada, and 11 kilometers northwest of Virginia City and the famous Comstock Lode. The principal springs and fumaroles occur in an area of about 2 ½ square kilometers, at altitudes ranging from 1,400 to 1,500 meters above sea level. Steamboat Creek, a north-flowing tributary of the Truckee River, flows along the east base of the principal spring terrace.

The climate is semiarid and is strongly influenced by the Sierra Nevada. The annual precipitation is about 30 centimeters at the springs but increases westward to about 75 centimeters at the base of the Carson Range, a branch of the Sierra Nevada.

Generally no more than about 12 springs have appreciable discharges at any one time, and these range from 1 to 75 liters per minute, with a total measured discharge of 150 to 375 l.p.m. The total flow of thermal saline water from the area is probably more than 3,000 l.p.m., including about 925 l.p.m. discharged directly into the bed of Steamboat Creek and an estimated average of 1,850 l.p.m. from the producing thermal wells.

In surface temperature, the springs range from about 45° to 95 ½° C., the approximate boiling point of pure water at the altitude of the area.

Geology

The springs issue from the northeast base of Steamboat Hills, which lie near the center of a north-trending structural trough bounded on the east by the Virginia Range and on the west by the Carson Range, a branch of the Sierra Nevada. The oldest rocks of the region are metamorphosed volcanic and sedimentary strata, intruded by granodiorite related to the Sierra Nevada batholith. These rocks are overlain in many places by Tertiary andesite, but the northeastern part of Steamboat Hills consists of granodiorite covered by a younger basalt flow, and the andesite is missing. The youngest volcanic rocks of the region are extrusive domes of pumiceous rhyolite of Quaternary age. Alluvium and hot-spring sinter probably range in age from earlier than the rhyolite to the present.

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(1) Published by permission of the Director, U.S. Geological Survey.

(2) Stearns, NORAH D., Stearns, HAROLD T., and Waring, GERALD A., Thermal springs in the United States U. S. Geol. Survey Water Supply Paper 679-B, 1937.

(3) BECKER, G.F., Geology of the quicksilver deposits of the Pacific Slope : U.S. Geol. Survey Mon. 13, pp. 331-353, 1888.

(4) BRANNOCK, W.W., FIX, PHILLIP F., GIANELLA, VINCENT P. and WHITE, DONALD E., Preliminary geochemical results at Steamboat Springs, Nevada. (In press, Trans. American Geophysical Union).

The dominant structures of the region are north-striking faults that to a large extent define the ranges. Parallel faults occur in alluvium of the trough and in Steamboat Hills. The fault zone localizing the springs is the part of this system that lies nearest the center of the trough. However, this zone appears at least in part to be a thrust fault dipping westward and is probably Quaternary in age. Because of its topographic position and a relatively high water table, the fault zone is a favorable route of exit for subsurface water.

Source of heat

The individual problems of the sources of heat, water supply, and mineral content are closely related, and each has a direct bearing on the others.

The most likely sources of heat for thermal springs are generally considered to be (1) : a) from the « normal » increase in temperature of the earth with depth, b) from an underlying body of hot, possibly molten, rock, c) from zones where there has been faulting of the rocks with resultant development of heat, d) from chemical reactions beneath the surface, or e) from the energy derived by disintegration of radioactive elements.

According to the theory of deriving heat from the « normal » or average regional increase in temperature with depth, the heat is transferred to the surface through the agency of deeply-circulating meteoric water. Van Orstrand (2) has shown that the average temperature gradient (^{o} C.–meter) in the sedimentary areas of the United States is almost certainly less than 1^e C. per 18,3 meters of depth and possibly as little as 1^e C. per 33,5 meters.

Proponents of this « normal » geothermal gradient theory have commonly assumed that the water arrives at the surface with little or no decrease in temperature from its maximum, and that the depth of circulation is indicated by the temperature. Within such a spring system from the recharge area to the base of circulation, the downward movement of cold meteoric water must lower ground temperatures. The temperature gradient, therefore, must be less than that of the surrounding « normal » region. Because water at the base of circulation may never reach the temperature of the surrounding rocks, and because it reaches the surface at an appreciably higher temperature than that of the surrounding « normal » ground, the thermal gradient within this part of the spring system must also be less than that of the « normal » area outside the system. It must be concluded, therefore, that an outstanding characteristic of a system of thermal springs deriving its heat entirely from a « normal » temperature increase with depth is a lower gradient than that of the surrounding region.

The temperature relations existing in a region where local excess heat is available from molten or other hot volcanic rocks is very different from that described above, and, in general, the temperature increase with depth is higher than « normal ».

Heat derived from friction due to movement of rocks on faults may be an important local or auxiliary source for thermal springs. However, in the western United States where numerous recent faults are known, there is little or no apparent correlation between the recency of movement and abundance or temperature of springs. The common association of thermal springs and faults is best explained on the basis of structural control rather than a genetic heat supply.

Day and Allen (3) and others have generally concluded that known chemical reactions are not an adequate source for the heat of thermal springs. A number of investigators have found no apparent relation between the temperatures of springs and the degree of radioactivity of the issuing waters. Atomic disintegration, therefore, does not appear to be an important direct source of heat for thermal springs, although indirectly it may be a critical factor in maintaining or supplementing the internal heat of the earth, and it may be one of the basic causes of volcanism.

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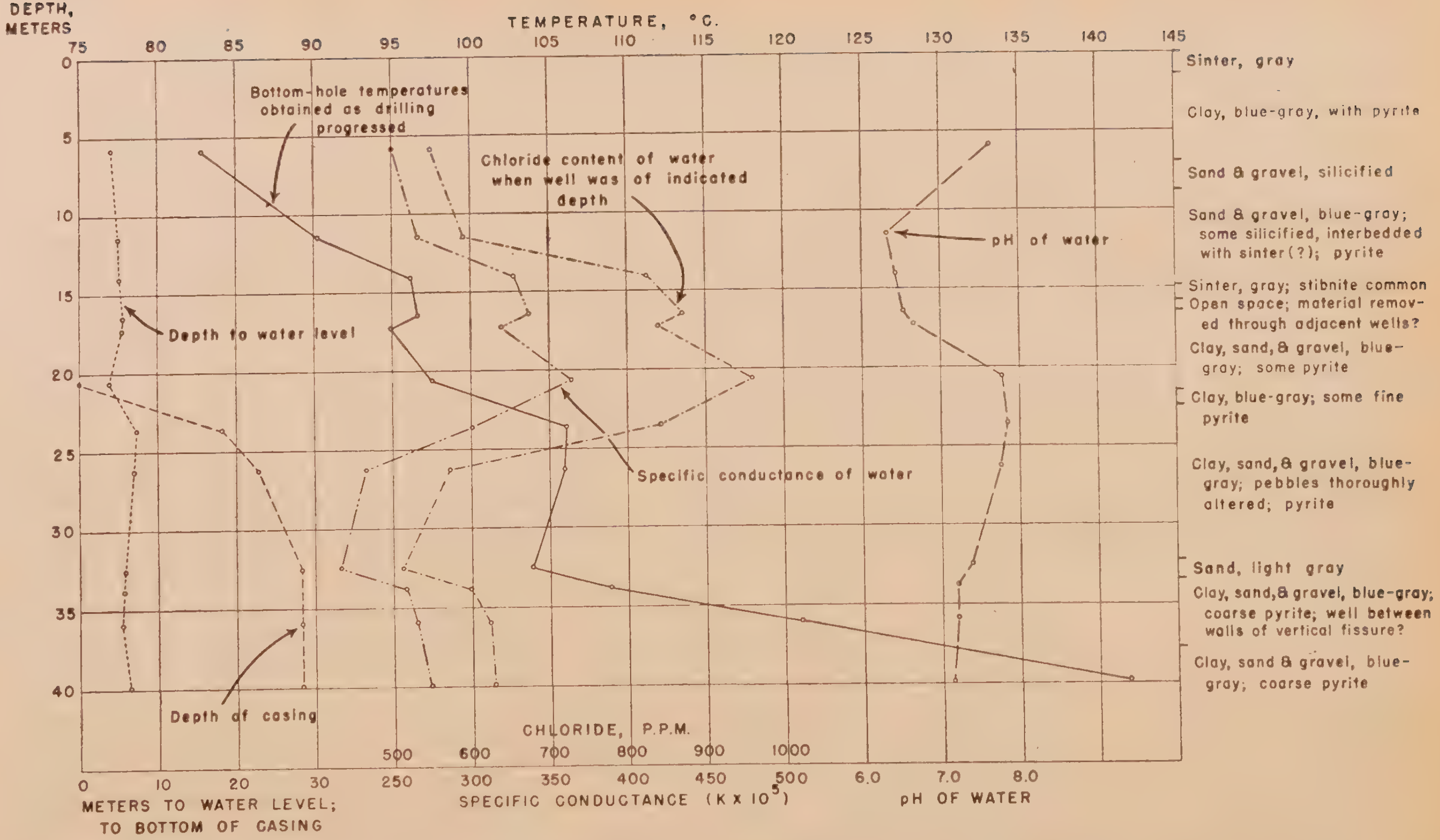
(1) Stearns, et al., *op. cit.*, p. 68.

(2) Van Orstrand, C.E., Observed temperatures in the earth's crust, in Physics of the earth, Pt. 7. Internal constitution of the earth, p. 147, McGrfaw-Hill Book Co., Inc., 1939.

(3) Day A.L., and Allen E.T., The sources of the heat and the source of water in the hot springs of the Lassen National Park : *Jour. Geology*, vol. 32, p. 183, 1924.



TEMPERATURES, CHLORIDE CONTENT, AND OTHER DATA OF STEAMBOAT WELL NO. 4, STEAMBOAT SPRINGS, NEVADA

Temperature gradient at Steamboat Springs

Measurements made in thermal wells at Steamboat Springs show very high temperatures at shallow depths. For exemple, the temperature at the bottom of a well 13.2 meters deep is 114° C., and that at the bottom of another 47.5 meters deep is 138° C.

Steamboat well N° 4 was drilled in 1947 in the southeastern part of the thermal area (see fig. 1). The rate of temperature increase is very irregular but within the explored zone the average is extremely high. The gradient can be computed by several different methods, depending on the assumed temperature at the surface, but none is completely satisfactory. The results range from 1° C. per 0.18 to 0.46 meter of depth, or at least 40 times the « normal » increase. The high temperature gradient at Steamboat Springs is believed to be strong evidence that the springs derive their heat from a volcanic source.

Composition of the gases

The gases evolved from the springs, excluding water vapor, contain 95 to 99 percent CO_2, and 0.1 to 1 percent H_2S; the remainder consists of nitrogen, argon, and oxygen. The « atmospheric » gases are probably derived in large part from the atmosphere and carried underground in solution in meteoric water, although there is little doubt that some nitrogen and argon is from a volcanic source. The gases are similar in many respects to those of other high-temperature springs and to volcanic emanations.

Composition of water

A typical analysis of water from Steamboat Springs is compared with those of other thermal waters of the United States (see table 1). The first five analysis are believed to be typical of most thermal water discharged at or near, the boiling point. In these, sodium and potassium are generally the dominant cations. The most abundant anion is generally chloride but this may be exceeded by carbonate, bicarbonate or, rarely, borate. Relatively high chlorine, boron, and sulfur, and a great dominance of total CO_2 (as carbonate, bicarbonate, and CO_2 gas) over calcium and magnesium are considered by the authors to be the outstanding chemical characteristics of high-temperature thermal springs, except in acid sulfate springs. The common occurrence of these same components in volcanic fumaroles indicates that they are at least in part of volcanic origin, although supplementary sources such as saline deposits must always be considered.

Spring N° 33 of Norris Basin, Yellowstone Park, is an example of the acid sulfate type, which has received much attention from Allen and Day (1). The origin of this type of water has not yet been demonstrated conclusively. The springs are generally situated on relatively high ground and are characterized by meager discharges in comparison with the chloride springs. The present authors believe that the majority if not all, of such springs are a superficial phase of thermal activity, generally occurring where the water table of saline spring water is below rather than at the surface. Steam, it is believed, is evolved by boiling from the water table and is condensed at or near the surface when it comes in contact with cooler ground or cooled, previously condensed water. In general the dissolved salts are left behind in the saline water but some intermixture may occur. The acid sulfate characteristics of the condensed steam are the result of atmospheric oxidation of H_2S that was evolved with the steam. It should also be mentioned here that where the supply of H_2S is abundant and discharge from the area is relatively low, the saline water itself can become acid and heavily charged with sulfate as a result of the same reaction.

Warm Springs, Georgia, is one of the clearest examples of a thermal spring that probably owes its temperature to a deep circulation of meteoric water. Even here there may be a remote connection with volcanism, possibly in providing a local higher-than-« normal » temperature gradient, but evidence concerning the gradient is lacking. It is clear, however, that there is no significant contribution

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(1) ALLEN E.T., and DAY A.L., Hot springs of the Yellowstone Natoinal Park: Carnegie Inst. Washington Pub. 466, pp. 65, 393-448, 1935.

Table 1. — Water analysis from Steamboat Springs, compared with other thermal springs of the United States (in parts per million).

Spring	Nº 27, Steamboat Springs (1)	Sulphur Bank, Calif. (2)	Geyser, Beowawe, Nevada (3)	Old Faithful, Yellowstone (4)	Iris Sp., Norris, Yellowstone (5)	Nº 33 Norris, Yellowstone (5)	Warm Springs, Georgia (6)
H	none	—	—	—	0.72	0.60	—
Ca	12	56.3	2	4	4	4	21
Mg	<1	57.1	none	—	2	2	12
Na	707	1389	239	372	336	48	1.6
K	75	68.1	33	31	91	30	3.6
Li	7	—	—	—	—	—	—
Al	<1	—	tr. (?)	none	3	1	—
Fe	<0.1	tr.	tr.	tr.	2	3	0.01
NH_4	<1	543	4	—	2	342	—
CO_3	20	none	173	100	none	none	—
HCO_3	292	1891	129	none	none	none	118
SO_4	129	1984	97	21	96	1105	7.3
S_2O_3	—	—	1	tr.	none	none	—
Cl	949	489	47	435	568	5	1.8
F	2.2	—	11	22	—	—	0.1
AsO_4	2.5	none	none	—	—	0.9	—
SbO_3	0.5	none (7)	—	—	—	—	—
NO_3	none	—	—	—	—	—	0.10
B_2O_3	192	1700 (8)	23	18	26	8	<0.1
SiO_2	317	67.4	449	357	476	303	23
Total	2705	8245	1208	1360	1607	1853	189
Temp. ° C.	95.6	80.6	boiling	boiling	boiling	hot	31.1
pH	8.2	—	—	—	—	—	—

from a volcanic source in the water of Warm Springs. The content of chloride shows no change from that of surface waters of the region, and boron has not been detected in either type. Also, the excess of CO_2 over the amount required for calcium and magnesium carbonates is approximately the same in both waters.

At Steamboat Springs the only available source for most of the Cl, S, B, and excess CO_2 appears to be volcanic. The bedrocks are notably low in carbonate and are not known to include saline deposits within a radius of many miles, except, for anhydrite and gypsum 18 kilometers to the southeast. The valley fill of Truckee Meadows north of the springs could conceivably contain saline deposits because of the grabenlike nature of the basin. This seems highly unlikely, however, because the Truckee River has probably always provided external drainage and because nonthermal ground water of the basin contains little if any more chloride than the surface water. In addition, the springs issue from granodiorite and alluvium near the edge of the valley fill. It is therefore highly improbable that any of the mineral content of the springs was derived from saline deposits or that appreciable CO_2 was obtained by reaction with carbonate rocks.

(1) Analyst, W.W. Brannock.

(2) Not previously published. Collected by C. P. Ross, 1938, and analyzed by M. Fleischer of U. S. Geological Survey.

(3) Allen E.T., and Day A.L., *op. cit.*, p. 174. Analyst, E.T. Allen.

(4) Idem., p. 249. Analyst, E.T. Allen.

(5) Idem., p. 469. Analyst, E.T. Allen.

(6) Hewett, D.F., and Crickmay, G.W., The warm springs of Georgia : *U. S. Geol. Survey Water Supply Paper* 819, p. 17, 1937. Sample from (East source, collected 1933. Analyst W.L. Lamar.

(7) Stibnite and possibly cinnabar were being deposited from the water. Oral communication from M. Fleischer indicated that as much as 2 p.p.m of SbO_3 could have been present but not detected.

(8) Reported as B_4O_7, 529 p.p.m., and H_3BO_3, 2,177 p.p.m.

Rhyolite domes

The springs are on a line of possible structural weakness connecting three or more rhyolite domes that may have been extruded after the oldest hot spring sinter was deposited. A structural and temporal connection with volcalism is indicated, and all other available evidence consistently points to volcanic rocks as the source of heat and much of the gas and mineral content of the springs.

Sources of the water

Since June 1945, systematic measurements have been made of spring discharges, temperatures, and mineral content. The object of these measurements was to determine the effects of rainfall and other changes in available water supply. In general, measurements were made one or two times a week, but daily observations were made throughout most of the interval of November 4 to December 30, 1946 (see fig. 2). Results of similar measurements for the first 8 months of the investigation are described elsewhere (1). Figure 2 indicates that the springs are not influenced by a slight precipitation but that a heavy rain has marked effects. The fact that the response is almost immediate, increasing discharge and decreasing chloride concentration and temperature, indicates that the effects are largely superficial in character and are due to an intermixture of rain water and thermal saline water in the uppermost part of the spring system. Probably only a part of this rain water fals on the spring terrace; the balance is believed to come from the small drainage basin immediately to the west. The influence of a heavy rain on the discharge is largely terminated within a few days. The chloride curve, it is believed, reveals a deeper influence. The high point in concentration on December 4, 1946, follows the previuos rain by 11 days. The record of June 1945 to January 1946, previously referred to, contains similar anomalous changes in the chloride curve at intervals of about 8 or 10 days after the two major storms of the period. The general rise in chloride from November 23 to December 4 is due to a gradual flushing-out of rain water that fell directly on the thermal area. The authors believe that the increased supply of meteoric water in the recharge area following each rain was eventually effective in deeper parts of the spring system in increasing the proportion of meteoric water mixing with a highly saline water. The lag of 8 to 11 days includes not only the time necessary to make effective the change in mixture (due to slow transmission of hydrostatic pressure in sediments of low permeability) but also the time required for the new mixture to reach the surface.

The records show that the discharge of individual springs as well as that of the group as a whole are influenced by many factors. In addition to those already discussed, barometric pressure is a major short-term factor; the springs tend to behave as inverted water barometers with efficiencies ranging from O to nearly 100 percent. Influence of earth tides has been detected in some water-level fluctuations but this is a minor factor. One of the outstanding characteristics of most individual springs is their ever-changing behavior. Many have stopped flowing, and new as well as old vents have begun to flow. Most of these changes are probably due to the filling or clearing of subsurface channels, but in addition, some major fluctuations have occurred for which there is no apparent explanation. It is possible that the volume of volcanic water emitted from its source is variable rather than constant, as is generally assumed.

Three-dimensional picture of upper part of the spring system

In general the highest concentrations of chloride and probably of boron are found near the center of the area of active springs, where the temperatures are also generally higher. The picture in relation to depth remained obscure until several wells were drilled for hot water in 1946-47. Steamboat well N° 4 (see fig. 1) was drilled in the southeastern part of the thermal area, where very high temperatures were encountered at relatively shallow depths. The South Steamboat well lies 250 meters farther to the south near the extreme southern edge of the thermal area. The data from this well are shown in figure 3. The outstanding features are : 1) temperatures and geothermal gradients are low compared to Steamboat well

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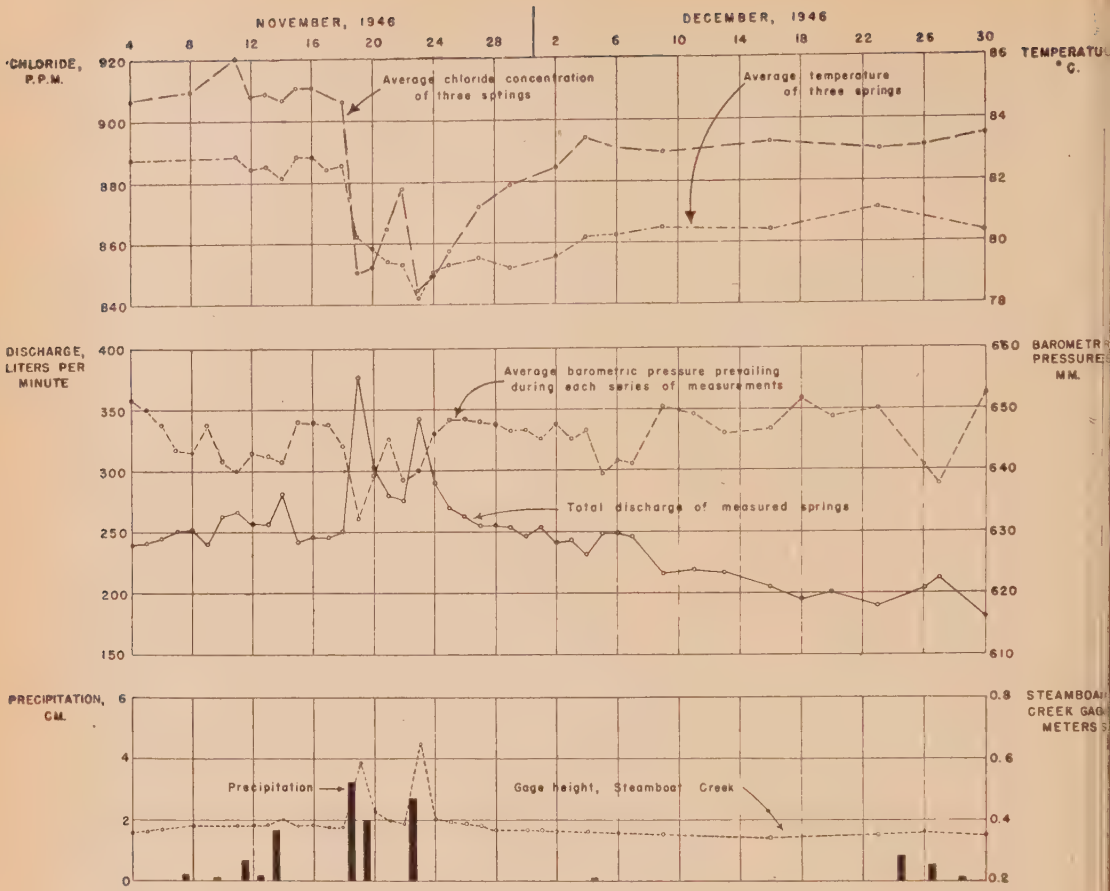
(1) Brannock W.W., et al., *op. cit.*

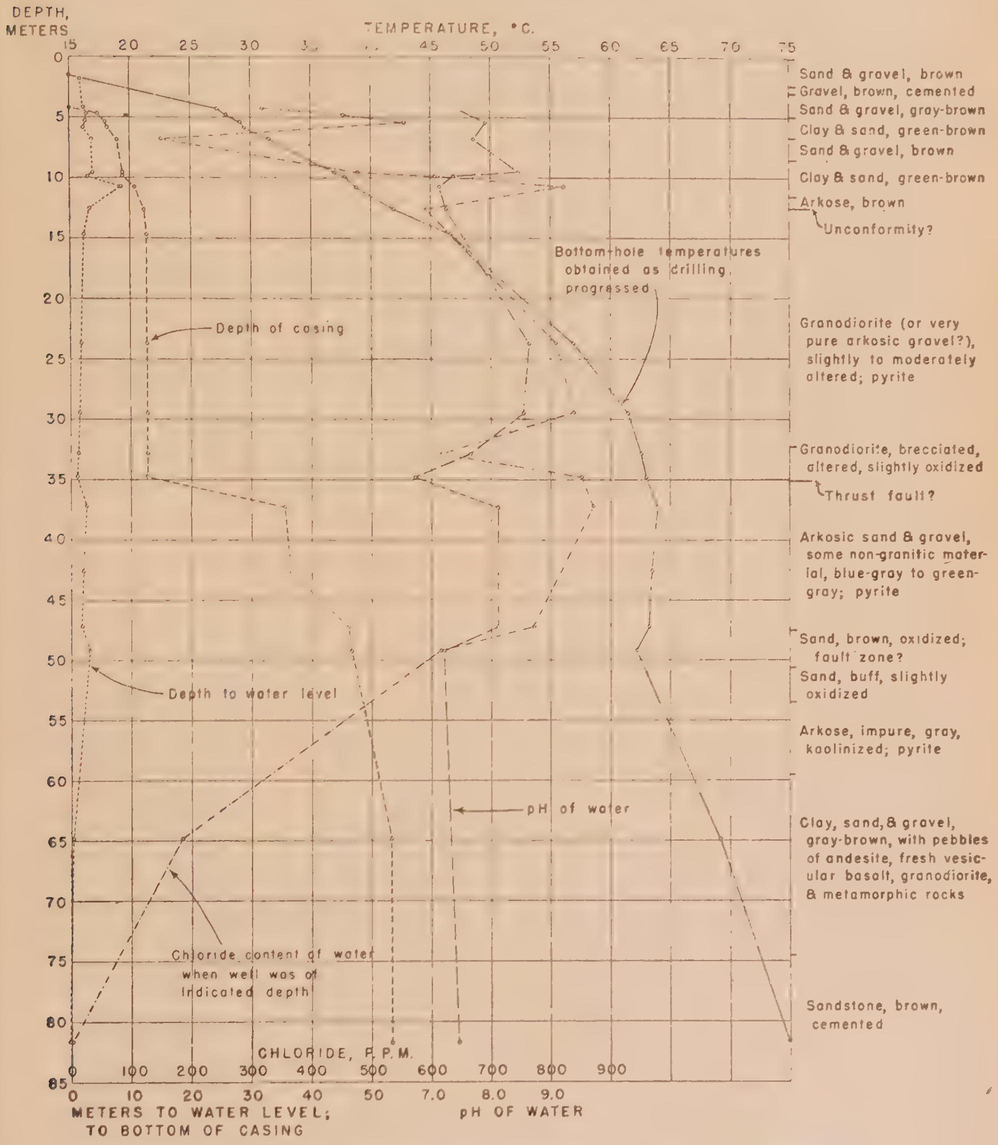


TOTAL MEASURED DISCHARGE FROM HOT SPRINGS COMPARED WITH RELATED FACTORS FOR PERIOD OF NOVEMBER 4 TO DECEMBER 30 1946. STEAMBOAT SPRINGS, NEVADA.

No. 4; 2) a major zone of saline water extends from the water level to a depth of about 50 meters below the surface; 3) horizons of less saline water occur within this major zone; and 4) water about below 38 meters decreases in salinity with depth. The well is cased to 53 meters, and water below this depth is almost free of dissolved mineral matter. Since the completion of the well, its chloride content has been about 8 p.p.m. For comparison, Steamboat Creek water contains 5 to 6 p.p.m. of chloride and average spring water about 900 p.p.m. The bottom zone in this well is therefore essentially water of meteoric origin. It should be mentioned in regard to figures 1 and 3, that observations and water samples were obtained from these wells at least 12 hours after the last previous drilling. The data are arbitrarily referred to the depth of the well during each series of measurements, although they are actually resultants of unknown combinations of conditions existing between the bottom of the casing and the bottom of the well.

The well penetrated sand and gravel from the surface to about 12½ meters and encountered what is believed to be granodiorite, in part thoroughly altered and shattered, from this depth to abouth 35 meters.

The remainder of the well is in arkose and cemented sandstone that is unlike any bedrock that crops out in the area. It is believed to be Quaternary valley fill, over which granodiorite has been thrust from the west, the fault being a part of the Steamboat Springs fault zone. If these relations are typical of the fringes of the spring system, they indicate a border zone strongly controlled by one or more faults. Near the surface, within at least a part of the border zone, saline spring water is migrating outward, roughly horizontally, and mixes with a small amount of less saline water that may be moving inward. Below this, a major zone of essentially meteoric water is migrating into the spring system and is being heated, in small part from above but largely from below. The movement of water is controlled by faults as well as permeable strata, and true artesian conditions are probably absent. The same fault or fault zone may be utilized by both the down-flowing meteoric water and the up-flowing hot saline mixture of meteoric and volcanic waters and is comparable in many ways to a huge convection system; the central part of the zone is extremely hot, so that the average density of the water is relatively



TEMPERATURES, CHLORIDE CONTENT, AND OTHER DATA OF SOUTH STEAMBOAT WELL, STEAMBOAT SPRINGS, NEVADA low in spite of its salinity; laterally from the central core but still within the fault system and adjacent strata, meteoric water is migrating inward and downward. Its average density is perhaps 5 to 10 percent greater than that of the saline water because of its low temperature. Such a system as this is adequate to explain the facts; artesian conditions, if present within the valley fill, would facilitate the underground movement of water, but they do not appear to be essential.

Fluctuations in the water level of the South Steamboat well, shown by records from a water-stage recorder, prove that the meteoric water of this well is definitely influenced by the saline Steamboat wells. When producing, the latter may be considered as more or less continuously erupting geysers. When the Steamboat wells are shut down, the water level of the South Steamboat well responds in less than half an hour, and begins to rise. A reverse effect occurs when the Steamboat wells are again erupted into production.

If the three-dimensional picture of the system that has been described is correct, the saline water of the central core becomes more and more saline with depth, and temperatures are in general closely related to the salinity of the water (see figs. 1 and 3). The evidence supports the belief that a relatively small amount of highly saline water is being evolved from extremely hot volcanic material.

Sources of the mineral content

Most of the evidence bearing on the sources of the mineral content of the thermal water has already been presented. In these particular springs, all or most of the chlorine, boron, sulfur, CO_2, arsenic, antimony, and fluorine is believed to be of magmatic origin. The source of silica and the common metals is not as evident; some may be of volcanic origin and some derived by solution from rocks adjacent to the channels of migration. It is highly probable that, if the volcanic water was originally rich in the alkali metals, much of the potassium originally in the water has been substituted for sodium in the wall rocks by means of base exchange. The absence of ammonia is notable, in contrast to the abundant ammonia in some other springs. Its very sporadic occurrence in thermal water suggests that, when present, it may come from sedimentary rocks containing organic material and that, when absent, the waters do not traverse sediments with an appreciable organic content.

Application of principles in other areas

High-temperature springs occurring in areas of recent volcanism commonly have characteristics indicating a volcanic relationship. However, the situation in regard to low-and intermediate-temperature springs in areas without known volcanism is rarely so clear. Conceivably, the heat but none of the mineral content of some of these may be volcanic. In others, there may be a slight volcanic contribution of water and mineral content, but so diluted with meteoric water that it is difficult to recognize. What criteria are of greatest general value in determining the origin of such a thermal spring ?

The authors believe that the thermal gradient within the spring system, as compared to that of the surrounding region, promises to be a helpful criterion. Little attention has been focused on this subject, and the data are meager. Even the small amount of available information is restricted to shallow depths in a very few areas. It now appears that thermal springs owing their heat to a « normal » temperature increase with depth should have subnormal gradients within the spring system, and that springs with close volcanic connections do have abnormally high temperature gradients. Springs remotely related to volcanism may have gradients within their systems that are no higher than so-called « normal » gradients.

A direct volcanic contribution of water and mineral content is most likely to be indicated by higher chlorine, boron, sulfur, and CO_2 than can otherwise be explained on the basis of composition of the meteoric water and bedrocks of the region. In areas where saline deposits are known and temperatures of springs are low to moderate, composition of the water is of little value although comparison of proportions of individual constituents may be suggestive.

Other criteria, such as composition of the gases, are probably of little value in moderate-temperature springs where there may be little or no volcanic contribution of water and mineral content. It is clear that in each case the problem is difficult and can be solved only by utilizing all the available evidence.

9) THE ARTESIAN WELLS OF THE HUNGARIAN BASIN

by

J. SÜMEGHY
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The water of the by ground-water supplied wells — in the Hungarian basis-system — is for drinking not recommendable. In most villages and other localities there is no sewerage. Thus all dirty-water of the lavatories easily percolates — through the very thin, very little filtering, and thus very little cleaning covering strata into the ground water, the level of which in the great plains is very high, and their motion and self-cleaning very little. The quantity of impurities in these ground-waters is extremly high,their NO_3 content-for instance-is often about 500-1500 mg. /ltr.

These circumstances forced the people of the Hungarian Basin to bore artesian wells, the first of which has been set up at 1830. Their number now is about 15.000-16.000. Most of them are on the deep, plain parts of the basins, and only a few on the marginal territories. Their convenient territory slowly decreases. In the years about 1900, for instance, the water of the — most general —150-200 m. deep wells rose higher than 100-102 meters over sea level. But now they only rise to 94-95 m.-s, which fact means, that they are only on these, or deeper lying territories positiv. The wells of 20-100 m. depth became slowly also all negativ. Generally, with the increasing of the number of the wells, their depth — in the same proportion — also increased, because always deeper and deeper water containing strata must have been reached, to get positiv wells.

This fact has many causes, one is — for instance — the hydrostatical pressure-decrease, as consequence of unrational water economy; there are many others too; but the explication can be found most easily by studying the geological structure of the Hungarian Basin, therefore, first I'll give a brief summary about it.

The series of strata in the Hungarian Basin — on some points till the basement — are known by exploratory drillings, set up during the lest years. According to these results, the Hungarian basin-system — between the Carpatians and Dinarides — has been formed of several crystalline and mesosoic mountain-ranges, and of more little interlying basins, filled and covered entirely or nearly entirely with Tertiary and Quaternary sediments. The geological evolution and structure of the single geological units is independent and different. Generally however, the whole basin-system, lying between the Carpathians, Alps and Dinarides, is a « centra mass » (median-mass, Zentralmassiv, Zwischenmassiv) with typical features of oratogen tectonism. The most important point is the southwestern corner of the Hungarian Basin, because this is the separating point of the Carpathians and Dinarides, and from this point are the several — above mentioned — mountain ranges diverging latitudinally in north-western, western and south-western directions.

The most important partly visible, ôartly underground -sunken-ranges, having parallel strike with the Carpathians, are : the Lajta mts. -Kiskàrpàtok range, separating the Vienna-Basin from the West-Hungarian one; the « Szombathely »-hills -Kőszeg mts.- Rosalia mts. range, separating the Bay of Graz from the West-Hungarian Basin, and the Bacher mts — the sunken crystallin basement near Mosonszentpéter — Inovec mts. range.

The Râba — Vàg — Nyitra ditch, and the row of the Turco, Lipto and Rajec basins are also parallel with the general strike of the Carpathians. All these above mentioned formations are on the marginal territories of the Hungarian-Basin.

The following formations — on the other hand — are rather inside the Hungarian « Central-mass », with various strike-directions : the Vots — Matzel — West-Central-Mountains range, jointing with the inside zone of the Carpathians; the Nagykanizsa — Bize — Nagyperkâta — Zagyva — Tisza ditch; the Orlica — Sleme — Kalnik mts. in connection with the underground structure near Inke, and with other sunken cristallin and mesosoic ranges of Transsilvania and the northern part of the « Alföld » (Great Hungarian Plain), and further in connection with the Radna mts. and the crystallin zone of the Eastern-Carpathians; the Er — Körôs — Mecsek ditch; the paleosoic range between the Sleme and Garics mts.; the « insulated » Baranya mts.; several crystallin and mesosoic underground ranges below the centre of the Alföld; and finally the ranges of the Hegyes and Bihar mts.

Formations of parallel strike with the Dinarides, are : the Drava and Sava ditches, the Bilo, Papuk and Fruska-Gora mts., and several crystallin and mesosoic ranges on and under the southern part of the Alföld.

This whole system, consisting of the above mentioned crystallin and mesosoic ranges and interlying ditches and basins, is surrounded with the young Carpathian and Dinaride mountain ranges. These two ranges are uniting themselves in the Balkan mts. Here the Alpine-type folding is already not intensiv at all, and therefore this region rather can be regarded as the foot-region of the orogenism.

Most of the above mentioned ranges after the Cretaceous period submerged, making thereby place (room) for the single parts of the Hungarian Basin-system : the West-H.-Basin with the Basins of Graz and Vienna, the Alföld, and the Transylvanian-Basin. But several blocks remained on the surface, between which, or separated by which these basins are lying. The basins are filled with Tertiary and Quaternary sediments. Amoung these sediments are the *eocene* and *oligocene* carbonaceous marls, flysses, limestones and clays very restricted yet. They can be found only on the sides of the Western-Central-Mts., in the Transylvanian Basin and in the Sava ditch in greater quantities. The late-Tertiary flyss-sandstone appearing near Debrecen, is supposed to be the prolongation of the flyss of the Radna mots. (1)

Miocene sediments are not so rare already, because the miocene transgression flooded great territories in the H. /Hungarian /Basin. / Mecsek, northern part of the west-H.-Basin, Transylvanian Basin etc. /. These sediments are mostly marls, clays, flysses, sometimes limestones and other sandy, loose sediments.

But the final formation of the H.-Basin-system took place only in the *pliocene*. In this period did the above mentioned several crystallin and mesosoic ranges entirely submerge, and in this period received the basins their final form. Characteristic sediments are marls interbedded with sandstones and clays. On several places these typical pliocene (pannonian) strata further submerged (west-H.-Basin, northern margin of the Alföld, Drava and Sava ditches, southern parts of the Transylvanian Basin etc.) and formed thereby little « cauldron »-like deepenings, and « depressions ». The upper-pliocene (levanthinian) strata, which are filling up these deepenings are mostly of loose, sandy character.

In the Quaternary, the sedimentation was of smaller scale, but the sinking of the basins continued.

The thickness of the Tertiary and Quaternary sediments in the basins is very different. The late-Tertiary sediments are generally on the margins, whilst the younger-Tertiary and Quaternary ones on the Basin-centres the thickest. Their total thickness — in general — is about 1.500-2.500 m.-s, but in the deep ditches they can reach the thickness of even 4000 m.-s.

In this Basin-system the Tertiary orogenism caused fractures and faults of great scale. Although there are locally — especially in the marginal, intensly folded ranges — features of foldings to be found too, but generally however, our « insulated » and « Central » mountains, — especially there, where the Tertiary formations take part in the forming of even the basement itself — have the typical features of fractured and faulted tectonism.

By the sinking of the basins, the upfilling strata became fractured and faulted too. These faults and fractures cut the greatest part et the Quaternary sediments too. (T. Roth Károly : The Geology of Hungary, Pécs., 1929.)

By the above mentioned evolution such a hydrogeological basin (« vessel ») system has been formed, in which several basins (water containing vessels) of different height are in connection with each an other.

In this basin system the western (Bay of Graz and Vienna) and eastern (Transylvanian) basins are higher, than the central lying « Alföld » and « Kisalföld » basins The crystallin ranges, separating these basins are very deeply sunken and thus they don't hinder the free water percolation from the higher basins into the lower ones; the mesosoic ranges don't hinder the percolation at all. Thus the moving of the ground-water from the higher lying basins — and other similar formations — into the central, lower lying basins is entirely free, without any hindrance. The bottom of this unific water containing « vessel-system » is the surface of the crystallin, variscian basement.

The water-gathering territories of this system are the limestones, dolomites, loose sandstones, pabbles, sands, loesses and loose volcanic tuffs of the most different ages. Their total territory is about 80.000 km^2, fifth of the whole basin-systems territory.

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(1) Jaskó S. : *Erosion and sedimentation in the H. Basin during the kainozoic era.* — Földtani Kôzlöny., LXXVII. vol. Budapest, 1947.

The water gathering territories are divided into two great groups. In the first group there are the limestones and dolomites. These rocks are permeable and thus the water descends in them, till it reaches an impermeable stratum. This fact means, that these carbonatic rocks consist — vertically — of two main zones : the upper, in which the water slowly descends (descending zone), and the lower — touching beneath the impermeable stratum — in which the descending water is gathered (water containing zone —). The greatest water quantities of the Hungarian Basin system are in these mesosoic limestones and dolomites contained. On the surface of the impermeable basement the water can move in the direction of the greatest slope. The hydrogeological importance of these carbonatic « water-container »-s is very great.

(F. Horusitzky : A viz a föld belsejében. The water within the earth. Hidrologiai közlöny., vol. XXII. p. 123-144. Budapest. 1942.

In the second group there are all other permeable formations and sediments, which are in communication with the sediments basins. These formations have their connection with the above mentioned carbonatic blocks mostly along great fault and fracture surfaces.

The importance of the first group is greater. Although these carbonatic blocks are separated by crystallin ranges and thus their water can't communicate into each an other, but — and therefore are they important — they widely penetratting all neighbouring strata they can supply them with water, mostly of high pressure.

In the second group there are all other permeable sediments of the basins. But most of them are only of restricted extense and thus they are in water containing and transporting only of local importance. Only the pannonian (pliocene) strata are extending to the whole territory of the H. basin system. This pannon formation is about 1500 m.-s thick. But its lower half, the lower pannonian, is mostly impermeable, and so only its upper half is of importance. This loose, sandy upper pannonian is several hundred m.-s. thick, and is of great importance as supply base of the most artesian wells. Firstly therefore, because their water is under hydrostatical pressure, already, and secondly, because they are on great surfaces in connection with the water containing limestones and dolomites, from where their water quantity can be increased. The base of the upper-pannonian strata, and the impermeable « base » of the mesosoic, carbonatic blocks are about in the same level, in the deepness of about 1000 m.-s. This level is also the lowest limit of the artesian wells, because, the deeper wells are supplied mostly by the — along tectonical surfaces — ascending hot (juvenile) waters aleady.

But in the Hungarian Basin we find such places too, where the origin of water also by the classical artesian well theory can be explained. There are such points — for instance — where the wells receive their water from higher located water gathering territories, and thus they ascend, cause of their hydrostatical pressure on the base of the « communication-vessel » principle.

Such places are further the sunken, and with upperpliocene (levantinian), and pleistocene sediments up-filled little local basin-parts and other local « cauldron »-like depressions. Such formations can be found interbedded into the pannonian series of strata in the west Hungarian Basin, in the Drava and Sava ditches, on the northern margin of the Alföld, in Zagyva-Tisza corner, on the central and southern parts of the Alföld, and on the southern parts of the Transylvanian basin. These local depressions are with such river and lake sediments filled up, which consist mostly of loose sands and pabbles, and thus their water containing capacity is very great. Their general thickness is several hundred m.-s, but above great ditches, they can reach the thickness of 1000 m.-s. too. These depressions are the richest in water, because in their sediments mostly several detritus and talus accumulations take part too, with their greatest dip always towards the depressions center. Mostly all surrounding, not sunken, and thus higher remained formations around these depressions are dipping also towards the depressions centre, and thus they also increase the water gathering territory.

This is the explication of the fact, that there are such points, where in little depth nearly all sandy strata contain water in great quantities, whilst on other points artesian wells can not be set up at all. In the Hungarian Basin system the great water containing formations are these local pleistocene depressions, filled up with the above mentioned loose - sediments, mostly sands and pabbles.

The various water containing basins, depressions, basinparts, cauldrons etc.

are in many respects independent, but there is of course a general connection between them by the — along fracture lines — moving waters. Therefore our artesian waters are mixed; that supposition is very problematical, according to which our artesian waters, deriving from mostly entirely separeted water containing « vessel »-s (?), in their chemical composition (salts soluted from the containing rocks, various hydratation products, etc.) indicate, and characterise the rocks of those vessels, in which they are contained. (Szâdecky — Kardoss E. : A nagyalfôldi artézi vizek főtipusai, és azok szintjelző értéke. — Bânyâszati és Kohászati Lapok. vol. LXXV, 16. p. 305-308. Budapest 1941.)

The NaCl containing waters of the central basins contain more agressiv CO_2 than usual — especially in the pabble strata — and also more Fe in solution. The waters of the most usual — sand — strata are able to bring also much Fe in solution, and their other salt content is also sometimes very high. Exceptions are the waters of the holocene depression along the « Kôrôslköz »; they contain much organic impurities, because their « oxygen-need » (for oxydising them) is very great. Further exceptions are the waters of the depression Győr, because they contain neither agressiv CO_2, nore Fe-salts in greater quantity. The waters of the pleistocene upperpannonian strata contain less salts than the waters of the lower pannonian; near the basement these lower pannonian waters approach the composition of the salt waters of the oil-fields. (Szurovy G. : A nagyalfôldi ujabb mélyfurâsok hydrogeologiai eredményei. — Hidrologiai Közlöny. 1947. 1-4. p. 17-20. Budapest. 1947.)

The depth of the wells depends from the geological formation. The wells supplied by the pannonian strata are poor in water, and of short duration; their water quantity is about 50-100 lits. /sec. But the wells of the levantinian (pleistocene) strata are rich in water (150-200 lits /sec), and of long duration. Most rich in water of them are the waters of the Tisza-Maros corner. The 280 artesian wells of the town Szeged supply 72.000 lits. /sec.! The 650 wells of Hodmezôvâsârhely sypply 30.000 lit.–sec. Many of the wells has 1000lts. /sec. All the wells of the whole Hungarian Basin system togheter supply 1.000.000 lits /sec.!

Their temperature is generally high. The water of the 150-200 m.–s deep wells is about 18-20° C., that of the 300-350 m.–s deep ones about 25-26° C. The cause of this fact is, that the geothermical gradient of the central basins is smaller (18 m.-s) than general. (Sümeghy : Die geothermischen Gradienten des Alföld. A M. Kir. Földtani Intézet Evkönyve. vol. XXVIII. p. 275-370. Budapest, 1929). According to F. Horusiczky the condensation of the water-vapours can supply enough heat for this heating of the basin sediments. According to him, the — along fracture lines — ascending vapours heated the sediments, and therefore is the geothermical gradient so small. This supposition is problematical however.

The artesian wells of the Hungarian Basin system can be devided into 3 groups according to depth. In the first group are the 25-100 m.-s deep wells. Their number at the beginning of this century was very great, but since that their number decreases. Its cause was the unrational water-economy, because often one water containing stratum had to supply several hundred wells. In Versec — for instance — every house had its own, about 28-37 m.-s deep artesian well. These wells are supplied mostly by the water of the surface-near detritus and talus accumulations.

Most artesian wells get their water from sandy strata in the deepness of about 150-250 m.-s. These are the wells of the above mentioned pleistocene (levantinian) depressions. Although they are densely set up, their water quantity does not decrease.

In the third group are the 300-450 m.-s deep wells. These get their water mostly from the upper-pannonian strata. These wells are of smallest duration. The increasing of their number causes the decreasing of their water-quantity and pressure.

There are also some 500-650 m.-s deep wells, but their number is very small.

In the last years some deep-drillings had been set up, for petroleum, hydrocarbon, salt etc. searching, their depth was several thousand meters, but their water was not artesian one already, but hot (juvenile) waters. Their temperature was even 147 C. near the basement.

10)

ORIGIN OF THE « JUVENIL » SUBSTANCES OF THE THERMAL WATERS IN HUNGARY AND THEIR QUANTITY OF HEAT

by

TIBOR SZALAI, Budapest

First of all we must consider the tectonic building, as the position of the reservoirs as well as the way of the rising waters and the ascending of the waters of the artesian wells depend on the tectonic system. I have to point out that the water-reservoirs of the thermal waters are the mesozoic part-geosynclines and therefore I have to mention their position. At the egde of the Eastern Alps a mesozoic virgate system is developing, the parts of which — as *Szalai* already has refered to in his study (13) — extend to the territory of the Carpatian Basin like the fingers of a hand. In this territory we observe it the strike of the Bakony the most perfect mesozoic strata including the lower cretaceous. In the middle triassic developed in the North West Carpatians a part-geosyncline, which involves also formations, which refer to terrigen influences, Besides we observe this part-geosyncline in the Mecsek in the lower triassic, in the Villâny territory in the middle triassic, in the Krassószörény mountains, the Bükk and the Bihar territory in the lower triassic, and in the Nagyhagymâs and Persâny mountains in the lower lias.

The leading part of the part-geosyncline in the Bakony ends only in the cretaceous, and in the Carpatian Arch the permanent geosyncline only develops in the senonian resp. in the eocene (14). This chief geosyncline bifurcates and one part extends at the exterior Carpatians, the other one along the Bihar group.

I want to report that several borings drilled in the area of the Great Hungarian Plain touched mesozoic formations.

The age of the strata of the boring of Hajduszoboszló II between 1447.10-2032.00 m. is problematical, perhaps a triassic strata was discovered (9. p. 55). The ascending waters contain $1.10.10^{-6}$ millicurie radium emanation per liter.

The boring of Köröszszegapáti VI touched a rock at 1795.5 m. depth consisting of conglomerate, crystalline shists and crystalline limestone talus. In the North East Carpatians I watched triassic and jurassic crystalline limestones which lay on the crystalline shists or in some places are folded on them and so we may suppose, that the investigated crystalline limestone is perhaps mesozoic too.

The Bugyi I. boring touched limestone, which may be regarded as upper triassic in 232 m. depth. If we connect this occurrence with the formation — which may be also triassic — drilled in the transdanubian Igal area, we obtain a direction parallel with the direction of the part-geosyncline of the Bakony. In the strike of the mentioned occurrences lays the triassic stratum of the boring of Tard. We come to the same conclusion if we connect the mesozoic strata of the boring of Köröszszegapáti VI. and the borings of Tótkomlós. I suppose that these directions indicate the geosynclines submerged in the depth.

The boring of Tótkomlós II. touched probably triassic basic conglomerate at 1661.5 m. the boring III probably triassic dolomites and conglomerates at 1691.5 m., the boring IV. probably triassic dolomites and conglomerates at 1685-1707 m., the boring VI. probably triassic dolomites and clayslate at 1535.5 m. (*).

The situation of the mesozoic formations in the virgate system is uncertain in the Bükk, the mountains of Krassószörény, Persány, Nagyhagymás as well as in several places in Slovakia.

We can watch a part of the mentioned sediments of the part-geosynclines on the surface, but their greater part lies below the surface and these may be considered as the drains of the Carpatian Basin. These drains are split into blocks by fractures, the above-mentioned closely drilled borings of Tótkomlós explain the structure of the sous-soil, because they all touch the mesozoic in different depths in spite of the fact that they are all drilled more or less on the same level.

There also may lie older strata as granite and other rocks between the mesozoic

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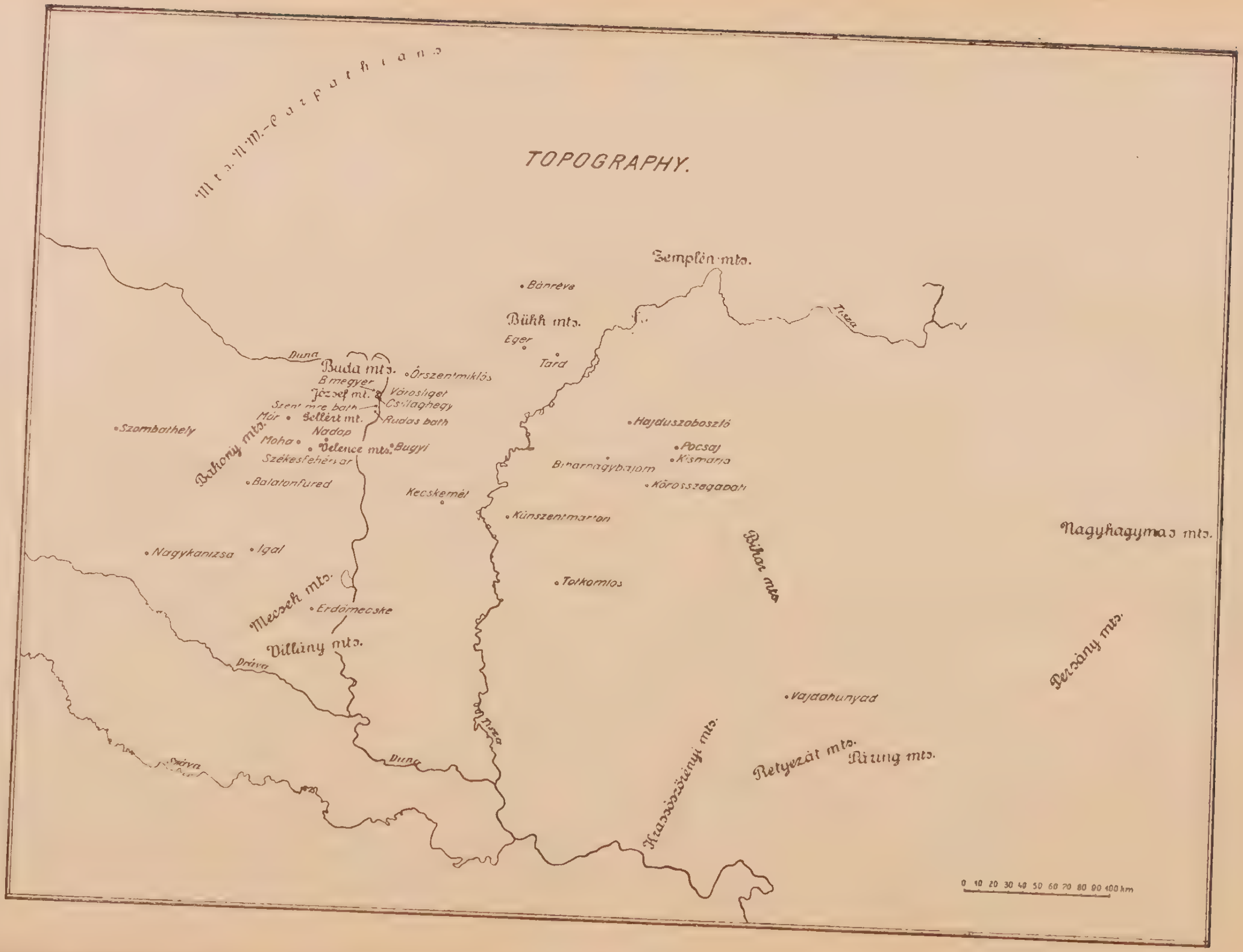
(*) I did not mention the borings in the northern border mountains of the Great Hungarian Plain, which lay upon triassic strata. (Budapest, Csillaghegy, Békàsmegyer, Orszentmiklós III., Tard.)



TOPOGRAPHY.

Mts. N.W.-Carpathians
Zemplén mts.
Bánréve
Bükk mts.
Eger
Tard
Tisza
Duna
Buda mts.
Örszentmiklós
B megyer
József mt.
Városliget
Csillaghegy
Szent Imre bath
Már
Gellért mt.
Rudas bath
Nadap
Moha
Velence mts.
Bugyi
Székesfehérvar
Szombathely
Bakony mts.
Balatonfured
Kecskemét
Hajduszoboszló
Pocsaj
Kismarja
Biharnagybajom
Körösszegapati
Künszentmarton
Tótkomlós
Nagykanizsa
Igal
Mecsek mts.
Erdömecske
Villány mts.
Dráva
Száva
Duna
Tisza
Bihar mts.
Nagyhagymas mts.
Perşány mts.
Vajdahunyad
Krassószörényi mts.
Retyezát mts.
Pàrung mts.
0 10 20 30 40 50 60 70 80 90 100 km

blocks, but we find them also in their foot walls, for instance : in the boring I. of Kismarja gneiss in the depth of 1551.5 m., boring II. in the depth of 1639 phyllite and micaceous shist, boring I. of Biharnagybajom, gneiss talus at 1003-1008 m., boring II dark grey chlorite shist at 1466.5 m. where we observe micaceous chlorite, biotite and rolled out quartz and felspar boring VII. a 1174 m. chlorite shists with quartz veins, boring II. of Körösszegapáti micaceous shist at 1531 m. boring III. amphibolite containing alkali felspar and plagioclase at 1842.5 m.; this boring lays open gneiss too, boring IV. tached amphibolite at 1858.5 m., boring V. amphibolite at 1630 m., boring VI amphibolite at 1469.7 m.. The drilling VII. discloses crystalline shists and the already mentioned crystalline limestones and boring VIII ground conglomerate and amphibolite talus at 1634.1 m. (*). These older rocks and especially the granites have-as we shall see later an important influence on the chemical construction, of the thermal waters and their quantity of heat.

The fracture tectonical construction is typical for the Carpatian Basin. We distinguish three chief tectonical unities in this territory. The central one is situated on the Great Hungarian Basin, the two other unities lie western and eastern from the Great Hungarian Basin. The part which lies on the Great Hungarian Basin is a paleozoic or praepaleozoic swell in the northern continuation of which we find a swell, quoted by *Sujkowski* (**) (11).

As to the part of the swell lying in the Carpatian Basin territory — *Szalai* informs us — that it may be observed in the northern direction far reaching in the region between the gulf of Botton and the Novaja Zemlja, in the South we may search for this swell in African areas and this swell is surrounded by two powerful transcontinental structures which cross each other in the Balkan area. The presence of the swell is based on the investigations of *Szalai* in the North-East Carpatians. Besides these facts and the arguments of *Sujkowski* I must point out that, up to now, unknown circumstance that the permian carbonic sediments of the Zempléni Sziget mountains differ entirely from the permian carbonic Bükk sediments, which are nearest to them so that we have to suppose the presence of the swell. As the mentioned formations of the Bükk mountains developed in a marine facies and as only carbonic layers of coal are known, in the Bükk mountains, there may be found antracite in several levels of the permian carbon of the Zempléni Sziget mountains. In this territory the whole sediment series refer to a litoral facies. The same facies is described by *Sujkowski* in the part of the swell lying North of this territory in the Pripet swamps and in Volhinia.

My following task is to explain the position of the waters in the water reservoirs. First of all we have to point out that these waters enclosed in the limestone, dolomite blocks are karst waters. We have to investigate whether these karst waters are to be found in one or in several levels and whether these karst waterlevels are connected in the different mountain territories. Are the waters supplied by rainfalls or are they lying depth since geological times. Before answering these questions I have to point out that the thermal waters occupy the reservoir rocks in several levels i.e. The mesozoic rocks contain several impervious slay rocks. In this way developes the upper karst water level which must be considered as deep lying ground water wherever the aquiferous rocks crop out. Under that ground water lies the level of the deep or the chief karst waters as *Horusitzky* quotes (4). Naturally it is possible too that newer investigations may find more levels than the two quoted ones. The karst waters occupy only the fissures and dissolved caverns along the tectonic planes in the limestones and dolomites. The deep karst waters touch the surface partly along the tectonic planes and eventuelly directly by the blocks coming out to day.

In some geosyncline territory the upper karst water levels can only develop uniform water levels in smaller unities as the water reservoirs are split into blocks. The deep karst water may eventually produce a united level in spite of the split blocks. *Szadeczky* quotes (12. p. 81) « We can explain the fact that the constituents of the triassic waters are nearly permanent in spite of the changing composition of the blocks by the circumstance that the waters of the various areas mix

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(*) In connection with the blocks which happened to be found in the borings, I should like to mention the research of Nopsca (6) of the territory of Hunyad : the big granite masses are everywhere older than praecarbonic chlorite shists which include granite inclusions of block size. The granite contains also gneiss, crystalline shists and amphibolitic gneiss inclusions.

The MASZOVOL was kind enough to give me all the data of the borings of the Great Hungarian Plain (exept those of Hajduszoboszlô II) I have to thank Mr Denisov who was kind enough to allow me to make use of the data and Mr L. Kōrôssy who collected them.

(**) Nopcsa writes : «There is a certain discordance between the chlorite shistes and the older crystallinic blocks in the Retyezàt and in a certain mesure also in the Pàreng areas. In these territories the older praeliassic (caledonic ?) horsts all develop SW-NE strikes.

up and are connected». Based on the study of *Raul Vajk* (15) we may also supose — if it is possible to follow the sediments of the ancient geosyncline till the Alps — that in that case several part-geosyncline karst water levels are connected. At any rate this supposition is likely in the case of the geosyncline of the Bakony. So there must be a subsurface river system in the depth where probably is only little water motion.

As to the level of the higher karst waters there is no doubt about the waters origin. It certainly is rain water. The question is more complicated if we consider the origin of the deeper karst waters : Are these waters accumulated since geological times or are they supplied by the rain water ? I am of the opinion that these deep karst waters accumulate through the infiltration of the rain water. To day most of the scientists do not deny the fact that the greatest part of the waters has infiltrated in the above described way, but some geologist believe that a part of the water, though not being juvenile, does not come from the surface but that it comes from the dehydralisation of the submerged blocks. Recently *Horusitzky* (4) studied this question. Generally, for my part, I do not deny the importance of the dehydralization, but I do mot think that it has an important part from the point of view of the « juvenile » constituents and quantity of heat of our hot water springs. As to our karst waters *Szadeczky* (12. p. 83) is of the same opinion and here I should like to mention that I have the same ideas regarding the artesian wells of the Great Hungarian Plain, Refering to *Haug* and *Van Hise* the lithosphere may have open fissures of twelwe km. *Adam's* experiments and *King's* calculations have proved that small splits can be open in the depth up to 18.000 m. Here I must point out refering to *Gautier* (2. p. 332) — that if granite and porphyr are heated up to 150-200 C° in a vacuum, the infiltrated water evaporates, the connate water freed in a vacuum after heating it up to 450-500 C°. The infiltrated water should be freed in the depth of 5000 m. and the connate water at 17.000 m. basing on a normal geothermical gradient, but we should reckon with the higher pressure and so the waters may be freed much deeper.

We must suppose that in those places where the dehydralisation of the blocks happened deeper than in the lowest limits of the open fissure systems, the freed waters certainly could not ascend without splits. Refering to the quoted calculations and theories, the dehydralisation of the connate waters may only happen in a depth more than 17.000 m. and so we can scarcely suppose that the deep karst waters were freed by dehydralisation. I believe that the dehydralisation has no importance neither in connection with the quantity of heat nor with the « juvenile » constituents, not even in case if the dehydralisation happens where the fissure systems are present. I.e. in the great depth there are impervious strata which make the eventual emanation of the vapors and gases more difficult. These preventing factors become the more important the greater the depth is. In case of freed water heat ascending through openings from very deep to higher levels also only small part of the water reaches the higher layer; the situation is not a better one than in a little depth. In case of natural springs the temperature of the water is lower than the temperature of the ascending waters of the borings of the neighbourhood. We may observe the same fact if the boring is not very deep so doubtlessly the temperature drops also within short distance as for istance in the Attila hot spring (Rudas bath). This artesian well bored to 36.8 m is the hottest spring of this artesian well group. Its temperature changes between 46,7° C and 47° C. The lowest temperature of the waters of this group is 35° C. An other example there is the 200 m. deep artesian well of Szent Imre bath the temperature of which is 50 C°. In this territory the temperature of the natural Mátyás springs changes between 36.3 C - 41.4 C. The relatively warm as well as the very hot spring ascend through drilled holes, for instance the two wells in the Városliget of Budapest; here the water temperature of the 970.48 m. deep wells varies from 73.4° - 73.8° C and those of 1257.10 m. depth have a temperature of 77° C.

The above mentioned train of ideas and examples prove that the waters developed probably by dehydralisation when ascending must lose much of their heat and naturally of their quantity. I therefore believe that it is inappropriate to state « that there are plenty of springs and wells filled with waters not of straight meteoric origin » (12. p. 77). This explanation seems still clumsier if we remember that the hot springs ascend constantly from the lower mediterran in the Budapest area refering to Schréter (10). In this case the waters might came to an end as for instance the oil or natural gas, but we have no proofs. that this may happen.

It is a question whether in case of our not accepting this kind of development where do the « juvenile » materials the radioactive substances, boracic acid, fluorit etc. and the quantity of heat come from and how do they get into the thermal Waters?

In 1947 the Hungarian Geological State Institute investigated radioactive materials under my directions but though we intended to pursue practical aims up to now we only got theoretical results. Our group of geologists and physicians investigated the granite of the Velence mountains in many places the Mecsek mountains (that is the unique occurrence of Erdőmecske) with the *Geiger-Müller* tube and they established that the effects are rather strong in this territory, on an average three times stronger than normal. *Sándor Szalay* (professor of Debrecen University) ascertained that his laboratory researches resulted in finding the occurrence of some ura nium (6-7.3 g. in a ton) and bigger quantities of thorium (40-73 g. a ton) in the blocks. In the granite of the Velence mountains the occurrence of fluorite is known for a considerable time (16). Recently *Főldvári* established the same occurrence in this territory in wider areas. Refering to the literature the boracic acid is a juvenile substance and *Longwell, Knopf, Flint* (5. p. 283) ascertain that the boracic acid proves the magmatic origin of the waters. *Vendl* (17. p. 11) is of the same opinion but *Goldschmidt* and *Peters* (3. p. 403) established from 14 german granite average samples, that they contain 0.001 % B_2O_3. Naturally these statements are relating also to the barite and the rite substances. Doubtlessly some substances were originally juvenile but now they are dissolved by the waters and so get into the springs.

The conclusion of these investigations is that the simplest way to explain the radioactive materials and other juvenile substances in our thermal waters is to suppose that the deep karst waters dissolve the substances from the granites wedged between the earlier geosynclines. Perhaps these waters also descend from granite or other primitive rocks which are the foot wall of the deep karst water level and there they dissolve the substances. Refering to *Weszelszky* (18. p. 30) the radium most probably emanates from the rocks underlying the dolomite. *Vendl* (17. p. 21) writes : « It is remarkable that in the waters of the two springs of Városliget which come from the greatest depth and are the hottest of all between the kation the equivalent percentage of the magnesium ion is smaller and the equivalent percentage of the natrium ion is greater and that between the anions the equivalent percentage of the hydrocarbonate ion is smaller as in the waters of the other springs. These differences shaw that the substances of the hot springs of the Vàrosliget do not come all from the dolomites but partly from the blocks under the dolomites. I prefer to sustain these theories because only a part of our thermal springs contain these substances, as for instance the spring groups of the Gellért mountain and the artesian well I of the Városliget, contain B_2O_3. But there is none of it in the artesian wells of the József mountain and in the artesian well of the Margit-Island (18. p. 29). We also find B_2O_3 in the boring of Pestszenterzsébet (9. p. 243), and in the boring ot Hadjduszoboszló Nu 1. (9. p. 216). The strongest radium emanation contain the waters of the springs of the Gellért mountain. Where the moving waters of the deep karst fissures can not dissolve the « juvenile » substances from the granite or other very old rocks, the springs do not contain them. Even if the origin of these substances could be explained by some valid theory in this case the occurrence of these substances should be found in bigger or smaller quantities in all thermal springs.

The problems of the quantity of heat production are closely connected with the above mentioned interpretations. I do not want to investigate the different theories which want to clear up the heat problems of the thermal waters but I want to continue this train of ideas I think of the radioactive heat and the radioactive heat production the atom spliting because the geothermic gradient essentially deviating from the normal in the Great Hungarian Basin territory may be explained by these. According to *Eligius Schmidt* (8a). the geothermic gradient is between 16-20 m., according to *Joseph Sümeghy* 18 m. on the average. The radium occurrence in the crust is of great importance to the thermal condition of the crust. Several geologists already worked an this problem, here I only want to mention the book of *Longwell, Knopf* and *Flint* (5. p. 278).

In the introduction of my study, I wanted to give a sketch of the structure of the Carpatian Basin. I described the virgation an the splited rocks, sunk in various depths which contain thermal waters and where granite blocks and other very old rocks are also wedget.

Where we observe geothermic gradients diverging considerably from the normal

I suppose that the water ascend near in the neighbouhood of the granite masses. I want to explain that the quantity of heat is produced by the heat energy through the atom splitting in the granites an this quantity of heat causes the considerable deviation of the geothermic gradient from the normal.

This quantity of heat influences the temperature of the whole basin submerged in the depth. I do not believe that the « juvenile » materials as well as the heat accumulation are essentially connected with the dehydralisation or the juvenile origin. And this means that there is no reason why to consider the origin of the thermal waters or an important part of them resulted by dehydralisation. We have to point out that the deeper karst waters are of meteoric origin and if there is any dehydralisation — which we can not prove — its part is of no importance.

From this point of view we may explain together with *Pálfy* (61) the fluctuation of the waters of the thermal springs with the different quantities of rainfall but there are also other important factors as the atmospheric pressures or the elevation or sinking of the water of the Danube, for instance in case of some thermal springs at Buda.

Basing on the above mentioned theories I have to suppose that in the depth of the Great Hungarian Basin there are considerable masses of radioactive material because the heat accumulation could only develop by them. As a supplement to my studies y want to mention that *T. Szelényi* adn *G. Csajághy* (14a) investigated 15 gas wells of the Carpatian Basin and in the water they found helium which can be expressed in hundredth resp. thousendth of percentage.

As the origin of helium on the earth is generally considered as the product of decomposition, we may also conclude from the presence of helium the occurrence of radioactive materials in several places of our territory.

It is rather interesting that the sample of the mentioned boring of Székesfehérvár taken from the depth of 820 ½ m. contains 1.54, the sample from the surface 0.0120 helium in vol. %, while the Szent István well of the Széchenу Bath (Budapest) contains 0.0267 of helium and the well of Elmond (Balatonfüred) 0.0137, the Agnes-well at Moha 0.0116 helium in vol. %. Among the investigated wells the above mentioned contain the biggest quantity of helium. It may not be ignored that the here mentioned occurrences lie in the same tectonical zone.

The already described swell must be considered as a tectonically strongly turned up territory, which had an independent part from the paleozoic in certain regards. This swell makes it possible that the radioactive materials, may come out to the day along the fissures resulted through the tectogenesis. Here I want to refer to the study of Sümeghy (10a) who ascertains that we find the lowest geothermic gradient along the fracture line of Pocs-Kunszentmárton. This fracture line lies in the strike of the Mecsek. In this facture belt are situated Biharnagybajom and Kismarja and in both places the borings investigated gneiss.

According to *Sümeghy* (10b) we find the richest natural gas boring along the fractures. In these wells there is besides the methane everywhere also carbonic acid.

E.R. Schmidt (8. p. 157) quotes : « that these two examples may prove that in case of the Great Hungarian Plan wells the calculation of the geothermic gradient is practically not influenced by the fall of the temperature caused by the gas blow out. The heat degree of springs which contain a great mass of soluble gases (f.i. CO_2) can be considerably reduced by the gas expansion ». *Therefore the natural gas and the small geothermic gradient do not depend on each other* and as we find both facts in this place we must suppose that in the fissures the low geothermic gradients are caused by the ascending heat, originated through the atom splitting. This heat develops in the already described way, and with regard to the above mentioned facts it is quite impossible to speak of heat originated by the migration of vapors blowing up from the depth.

I pointed out in one of my last studies (14) that structures caused by contraction are followed by dilatations. I supposed that during the contraction the earth a crust of various specific weight shrank in different ways, the exterior crust pressing the underlying one. These mechanical effects fractures and dilations, are caused by a reaction of the pressure, *but also in this case we must consider besides the mechanical reactions the expansive force of the atom splitting.* In our territory the pleistocense movements can be proved and the recent earthquakes (Szombathely, Mór, Kecskemét, Eger) are the proof that the equilibrium is still unstable.

Gárdonyi (2a) comparing the results of the investigations engeneers have made now and some times ago, states, that a zero line can be followed in the North-East of Nadap through Budapest-Bánréve, in the South-West in the direction of Nagykanizsa and in the Dráva valley. Western of the zero line we can observe the raising, and in the S-E direction the sinking of the territory.

As the result of my studies, the deviation of the geothermic gradient from the normal must be explained by the heat produced by the atom splitting and by the heat produced by mechanical action.

The difference of temperature of the thermal spring refer besides the above mentioned facts that the waters ascend from waterbearing blocks, which submerged in different depths. The thermal springs of Budapest are excellent examples, because ascending from lifted up strata their temperature is a lower one, but ascending from submerged blocks, their temperature is a higher one.

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11)

HYDROGEOLOGY OF BUDAPEST BITTER MINERAL WATER WELLS

by

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Budapest bitter mineral water wells are situated in the quarter of *Buda*, S and SW of the *St. Gellért Hill*, upon a low-lying plain built up of Middle Oligocene clay called regionally Kiscell Clay. The wells form three groups situated in the following circuits of the quarter at given elevations : *Làgymànyos Group* 105 m, *Ormező Group* between Ormezö and Dobogó Hills, 108 m, *Őrsőd Group*, between the last named and the Péter Hill, 110 m above sea. (*Fig.* 1).

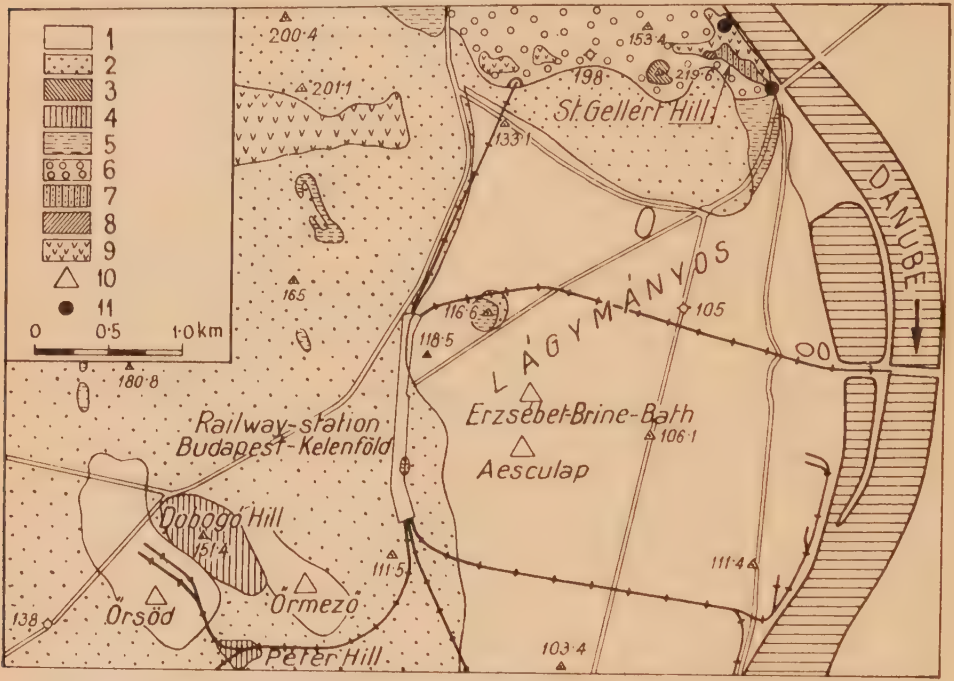

FIG. 1. — *Geological sketch of the Làgymànyos region of Budapest* (Map of *Schafarzik* revised).
1 = Bog-earth and fine sandy clay; 2 = Loess; 3 = Fresh water limestone; 4 = Upper Oligocene sandstone; 5 = Middle Oligocene Kiscell Clay; 6 = Lower Oligocene Buda Marl; 7 = Bryozoa Marl (Lower Oligocene); 8 = Nummulitic Limestone (Upper Eocene); 9 = Upper Trias dolomite; 10 = Strong bitter water-11 = Thermal wells. — Band between two pointed lines denotes Old Holocene Danube Riverbed.

The mineral water is formed and is mostly located in the *Kiscell Clay*, which contains little colloids and is sandy and muddy. Fattets clay contains less than 30 % « raw clay » according to screen analysis.

The clay with includes streaks of sand and sandstone has in its natural moist state, not exposed to weathering, bluish-gray colour, in dry state lighter gray. This colour is due to the presence of iron pyrites. Where the clay has long been exposed to the atmosphere the pyrite has altered to limonite and the less resisting other minerals have been more or less decomposed (1). Such clay assumes yellow or reddish-yellow shade. This observation has been proved by chemical analysis.

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(1) A. VENDL : Der Kisceller (Kleinzeller) Ton. — *Annales Inst. Reg. Hung. Geol.* XXIX 2. (The Kiscell Clay. Yearbook of the Hungarian Geol. Inst. V. XXIX. 2. *In German*).

The mineral water is formed in the same way in all three regions. The *Làgymànyos region* is however somewhat differently built up from a geological point of view.

The Làgymànyos region, an old Holocene Danube terrace, is a flat land gently sloping toward the present bed of the Danube River. The substratum of the alluvial deposits is Kiscell Clay, The clay is in its upper layer, mostly of 2-4 m thickness, yellow, deeper layers are blue. The surface of the clay is uneven, the River has washed off the sandy belts of the clay more than the fat portion which contains more colloids, as this has been proved by borings. In the Làgymànyos region, the greatest part of the water is located in the sands and gravels overlying the clay, but a smaller portion of it occurs in the clay itself.

Although the Kiscell clay contains in some places interstratifications of sand which are porous, in its great mass it is yet impermeable to water. However, in many places the surface of the clay was denuded by the Danube and it was temporarily flooded by high water. On other places it was exposed to the atmosphere, even though sometimes covered with thin layers of sand permeable to water and air. In such places, before it was covered by thick deposits, the clay had been exposed to the action of the atmosphere : its upper layers became less dense and its colloids partly coagulated. The texture of the clay became granular, and such clay may also contain ground water so long as this is not made impossible in virtue of its elevation above ground watei level. Loose, granular and permeable clay is not compressed even by later superincumbent strata of 5-10 m thickness so far as to loose its granular structure. Of course, more or less flowing ground water can also exist in sandy layers of the clay.

The location of ground-water depends in such places : 1) on the irregularities of the surface of the clay, 2) on the granular texture of its upper layer, 3) on the structure of the older Holocene Danube deposits. There is sufficient ground water present in the sufficiently deeply situated granular layer of the clay and above it where older Holocene deposits are enough permeable. There is no or very little flowing ground water where older Holocene deposits are impermeable or where the uneven surface of the clay lies at a level above that of the ground water corresponding to high water of the river.

The ground water level is variable in consequence of the variable tickness of the granular upper layer of the Kiscell clay and in consequence of its variable elevation : differences within some m. distance can amount to some dm. The groundwater surface generally dips toward the Danube, that is in the direction SE, The water-level of the Danube influence immediately the level of subsoil water.

In the Làgymànyos region the subsoil water is also found in interstratified coarse sands and fine gravels in the clay, but there are few data available thereof.

Essential active constituents od the Buda Bitter Mineral Water are sulphates of sodium and of magnesium. Of secondary importance are calcium sulphate and rock salt.

Geological conditions of the *Őrsőd bitter mineral water region* are somewhat different from those of the Làgymànyos. The plain valley of Őrsöd is built up of Kiscell Clay while the two bordering hills, Őrsöd and Dobogó. consist of Upper Oligocene sand and sandstone. (*Fig.* 2).

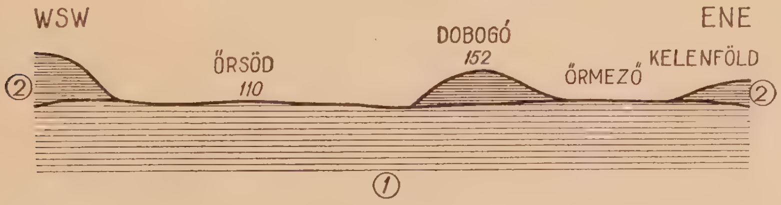

FIG. 2. — *Geological cross section of the Orsôd, Dobogó Ormezo bitter water fields.* (Taken along the strike). 1 = Kiscell Clay; 2 = Upper Oligocene sandstone.

These are pervious to water and lie in conformity upon the clay : the water percolating through the former hits an impervious layer on the surface of the clay.

The above named hills are remains of a formerly much more extended cover of Upper Oligocene rocks. From the point of view of water accumulation the loess is also of importance.

The Örsöd bitter mineral water region forms a flat basin the former Upper Oligocene cover of which has been washed off by the surface water, winding to and fro in consequence of the slight slope. This erosion had only been effective down to the high water level of the Older Holocene Danube. This level is so high in this region that the surface waters have hardly penetrated into it. On the NE and SE border of the valley thin fragments of the Upper Oligocene sand cover have been preserved here and there.

Kiscell Clay forms the bottom of the flat basin. There were marshes here before, precipitations are now drained off by means of an artificial canal toward the east.

The whole of this flat land of about 50 hectares is at about 100 m above sea, which is about the level of the Older Holocenic Danube. The Kiszell Clay of this region, only temporarily covered with water, was made porous on its surface by influence of the atmosphere. Its pyrite was oxydized and the top layer of the clay obtained a granular texture by partial coagulation of its colloids. This granular clay locates here the subsoil water. Fluviatile sand and gravel deposits, characteristic for the Làgymànyos region are not present here.

In place of periodic swamps, bog-earth has been formed which covers the granular clay in a thickness of some dm. The subsoil water of this very flat plain partly stagnates in the upper, yellow and granular layer of the clay, partly it percolates slowly toward the SE, to the Danube.

This upper, yellow layer of the clay, of 2-3 m tickness, is so filled up with bitter mineral water in winter and spring that the wells of 6-8 m depth almost overflow.

On the borders of the plain, the subsoil water percolating through the Upper Oligocene sandstone into the granular clay is only feebly bitter or even entirely fresh. The mineral water has its highest concentration in the middle of the valley where there is no Upper Oligocene cover; but even here great differences of concentration are observed.

The Mineral Bitter Water Region of Örmezö is also an alluvio-eluvial plain built up of Kiscell Clay. Thus the structure of all three regions is essentially similar : the Kiscell Clay is either exposed or it lies at very shallow depth under the surface.

The first acceptable explanation of the formation of bitter mineral water in these regions was given by Prof. J. Szabó (1).—Following results of my investigations confirm his theory.

In course of oxydation the sulphur content of the pyrite is oxydated to sulphuric acid, that is, to sulphate anion. The acid thus formed decomposes carbonates contained in the clay or hydrocarbonates — $Ca/HCO_3/_2$, $Mg/HCO_3/_2$ — dissolved in subsoil water and attacks also decomposable silicates, Sulphates, mainly those of Na, Mg, Ca and K. The major part of calcium sulphate cristalizes as gypsum.

Sulphuric acid thus formed acts upon silicate minerals contained in the clay; on the other hand, it is used up for further oxydation of ferro-sulphate, an intermediary product formed in the course of oxydation. Under the action of sulphuric acid, sodium and potassium sulphate is formed of the Na — and K — content of the clay : both readily soluble in subsoil water.

According to chemical analysis much more K than Na is contained in the Kiscell Clay, yet there are much more Na- than K-ions present in subsoil water. The major part of K is present as muscovitte in the clay, which resists to sulphuric acid and shows no change whatever under the microscope. From biotite and chlorite present in the clay, Epsom salt is formed.

The developing Mg-, Na-, K-, Al- and sulphate-ions pass into the subsoil water. Iron leached from biotite is oxydized and produces ferric hydroxide and sulphuric acid by hydrolysis. Ferric hydroxide is precipitated as limonite within the biotite crystals along their cleavage planes or outside of them. Chlorites are decomposed in the subsoil water. Limonite is also precipitated eventually within the chlorite leaflets themselves.

According to microscopic examinations some grains of glauconite are also decomposed in the clay, their exterior is either entirely yellow or at least spotted with yellow. Many of them contain irregularly shaped grains of limonite. From the decomposition of glauconite alcali-ions pass into the subsoil water.

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(1) J. Szabo : Geology of Environs of Budapest. Scientific Monography of the Capital, Vol. I, p. 1-14, Budapest 1899 (In Hungarian).

All the iron contained in pyrites or liberated by the decomposition of silicates is bound in the form of limonite.

To explain the presence of plagioclase in the Clay Szabó supposed that the Kiscell Clay should contain andesite tuff. Felspars are however, ingridients of the clay without any relation to tuffs. The plagioclases of the clay belong to the acid members of the species. In no mineral water well were found layers of andesite tuff.

The process of oxydation of pyrites is somewhat altered where traces of coal are present in the clay : in this case reduction also takes place. In presence of sufficient air however, oxydation goes on in this case as well and sulphates are formed.

The salt subsoil water coagulates the yellow clay in contact with it, renders it granular and pervious to water. This granular zone is more or less deep according to the degree of oxydation. Therefore the surface of the dense, blue and impervious clay is of variable elevation as well. The subsoil water is therefore most probably not continuous and its level is locally different. This is especially the case at low water level. The water level, of course, depends upon the level of the Danube as well. The greatest difference of level gauged on the same day at the Erzsébet-Brine-Bath was 2,2 m.

In the Őrsöd region the subsoil water has different levels as well; simultaneous measurements gave depths of 1,45 and 2,15 m. measured from the almost horizontal surface. Even in two contiguous wells differences of water level of 0,2-0,3 m. were observed.

For the formation of sulphates much air is necessary. In a newly dug well or ditch the accumulated water has initially a low sulphate content, which most substantially increases after the water has been exposed to the atmosphere for some days. The owners of the wells have long learned from experience that wells have to be sunk with a possibly large diameter, and their masonry has to be of such kind as allows of a free access of air or oxygen to the clay.

The sulphate content of the Kiscell clay is low in comparison to the great amount of subsoil water. However, under favourable hydro-geological circumstances, the solution can obtain locally a very high concentration. This concentration depends besides on the sulphate content of the clay also on the velocity of flow of the subsoil water and on its rate of evaporation. The velocity of flow is generally very low due to the feeble head of water in the flat regions, but is it not low enough to produce in itself a strong concentration of the salts. Within smaller areas however, the water moves extremely slowly, especially where it is located in the upper granular layer of the clay or partly in the fine clayey-muddy Danube deposit overlying the clay. In some places, at very low water level, sections of subsoil water of different extension may stagnate as subsoil ponds for some time in the hollows or valleys formed by the uneven surface of the clay. In such places the water is very much saturated with salts and becomes veritable bitter mineral water.

When the subsoil water level lies at shallow depth and loose sediments cover it, the rate of evaporation can be very great especially in the warm and dry seasons.

In the Làgymànyos region the salt content of the water is very variable due to the difference in local conditions, and such variations occur step by step as shown on the following table the numbering of which is identic with that of fig. 3.

From some data it is evident that the sulphate content has increased after considerable time (153, 154, 158, 161). Continuous supply of oxygen has furthered oxydation and more sulphate has been formed.

Results of analysis of the mineral water of the Erzsébet Brine Bath show the quantity of sulphate dissolved. The concentration of salts is very different here as well. Well Màtyàs II contains most 28.38 g and Ilona Well least 9.15 g SO_3 in 1,000 g water. The water of Aesculap Well contains 20.55 g SO_3/l.

The zone of highest concentration forms a belt of 300-400 m. width striking in SSE direction through the site of the Brine-Bath. The surface of this zone lies still somewhat deeper than the immediate surroundings. Borings show that this zone was one of the main branches of the Older Holocene Danube and was filled up with fine-grained, clayey and muddy sediments impervious or feebly pervious to water. Therefore the subsoil water percolates through the ancient alluvial beds and in the granular Kiscell Clay under them, very slowly and almost stagnates therein. It is enriched with salts especially where it occurs in small basins of the river arm. Of course, beyond the river bed mineral water occurs as well where there is much pyrite in the clay and conditions of slope and of evaporation are favourable.

SO$_3$-content of one liter mineral water in mg :

30. = 200 (1)
31. = 250
32. = 290
33. = 411.5
34. = 1140
35. = 674
36. = 328
37. = 922.9
38. = 342.8
39. = 1200
40. = 240
41. = 1708
42. = 767.1
43. = 28.8
44. = 17.9
45. = 389.9
46. = 220.8
47. = 691.4
48. = 3285.7
49. = 920
50. = 2247.7
51. = 2043 and 2990(2)
52. = 340.2
53. = 600.1
54. = 1033.9
55. = 867.5
56. = 945.8
57. = 1907.2
58. = 721
59. = 461.6
60. = 508.9
61. = 512.9
62. = 835
63. = 6749
64. = 496.3
65. = 514.1
66. = 833.8
67. = 1288
68. = 510
69. = 360.6
70. = 1769.1
71. = 6470
72. = 2300 and 5090
73. = 14034.2
74. = 303
75. = 16502
76. = 17280
77. = 555.3
78. = 785
79. = 445
80. = 586.7
81. = 590
82. = 140
83. = 570
84. = 400
85. = 128.3
86. = 1121.4
87. = 2147
88. = 451.9
89. = 1780
90. = 447.9
91. = 517
92. = 2013
93. = 633.3
94. = 316
95. = 852
96. = 419
97. = 931
98. = 1023
99. = 294.8
100. = 650
101. = 476
102. = 1050 and 1586.7
103. = 42
104. = 738
105. = 1556
106. = 1140
107. = 1331
108. = 676.2
109. = 4730
110. = 5260
111. = 1760
112. = 1541
113. = 779
114. = 780.4
115. = 1118.6
116. = 463
117. = 270
118. = 1774
119. = 976.4
120. = 2001.5
121. = 2276
122. = 3921
123. = 789.7
124. = 8048
125. = 1896.7
126. = 521.2
127. = 482.1
128. = 1586.7
129. = 9927
130. = 487.6
131. = 460.9
132. = 720.5
133. = 538.4
134. = 4322
135. = 7366,6354(3) 7809.2,8172
136. = 8624
137. = 1883
138. = 446.5
139. = 710
140. = 569.2
141. = 3576.6
142. = 4722
143. = 1345.5
144. = 1118.6
145. = 1260.5
146. = 1336
147. = 2509
148. = 9628
149. = 15002
150. = 12106
151. = 13377
152. = 9004.4
153. = 2538 later 6224
154. = 2513 later 2091
155. = 3212
156. = 73
157. = 780.4
158. = 1467.6 and 5734
159. = 7674
160. = 1083.6
161. = 809 later 1522 afterwards 2527
162. = 2621
163. = 1851
164. = 2147.9
165. = 1611.8
166. = 700.9
167. = 6249
168. = 1580.8
169. = 935.2
170. = 763.1
171. = 764.5
172. = 336
173. = 843.6
174. = 801
175. = 732.5
176. = 698.2
177. = 597.7
178. = 108
179. = 151.6
180. = 145
181. = 104.3
182. = 267.8
183. = 148.1
184. = 136.8
185. = 110
186. = 351
187. = 323
188. = 569.8
189. = 481
190. = 847.4
191. = 496.1
192. = 415
193. = 191.4
194. = 271.5
195. = 148.5
196. = 118.6
197. = 222.4
198. = 267.1
199. = 210.9
200. = 287.5
201. = 241.7
202. = 493.4
203. = 437.3
204. = 109.5
205. = 301
206. = 219.5
207. = 280.8
208. = 303.8
209. = 214
210. = 679.1
211. = 2866
212. = 1357.7
213. = 679.1
214. = 731.7
215. = 945.8
216. = 1010
217. = 2143.9
118. = 26018.1
219. = 591.2
220. = 1852.4
221. = 1613.4
222. = 377.3
223. = 516.1
224. = 1572.1
225. = 793
226. = 1067.2
227. = 248.9
228. = 97
229. 670.3
230. = 1003.9
231. = 94.5

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(1) Numbers, 30-34, 38, 40, 49, 51, 58, 71, 72, 74, 79, 81-84, 89, 94, 95, 97, 105, 109-113 ,116-118 and 13a are referred to 1 kg water.

(2) Two contiguous borings.

(3) After different examinations.

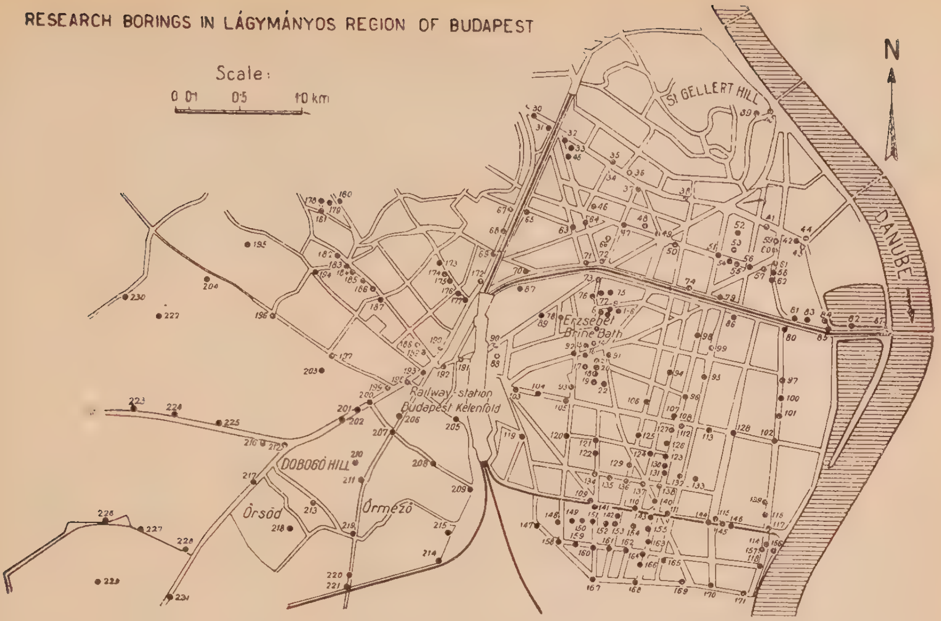

FIG. 3. — *Research borings in Lágymányos Region of Budapest.* N° 30 - 231 denote locations of holes.

Traces of another ancient river bed strike the eye E-of Fehérvàri Street in the form of a low-lying belt (1). The slope of this eastern river bed is greater than that of the Brine Barh. The subsoil water moves along it faster and does not reach the same concentration as in the former.

In the bitter mineral water fields of Örsöd and Örmező the subsoil water flows very slowly through the upper, yellow layer of the clay and the sandy layers locally interbedded with same, toward the Danube. The bog-earth covering the clay suffers cracks in summer across its full thickness : evaporation of subsoil water is furthered therewith. Slow flow of water and strong evaporation lead to strong concentration of salts in subsoil water.

In the plains of Örmező and Örsöd rainfall soaking through the Upper Oligocene sandstone has an important part in ground water collected in the Kiscell Clay. Especially in the Orsöd region it is striking that the ground water on the NE border of the plain is fresh because rainfall soaking through the Upper Oligocene sandstone still forms a very thin solution. In the interior of the basin the water is concentrated and turns highly bitter.

Bitter Mineral Water Wells. The Erzsébet Brine Bath in the Làgymànyos Region has had altogether 22 wells of 3.86-6.62 m depth and not all have reached the unaltered blue clay.

In June 1947 the Author intended to establish by borings whether further mineral water wells could be dug on the estate. It was found that the sulphate content of the water was very variable in the borehole already : SO_3 content was 362-5640 g/l.

To determine the sulphate content water samples were taken during drilling. Repeted sample taken after 5 days showed that the sulphate content had increased to ten times its former value.

Other tests have shown that the sulphate content can decrease and even disappear under certain circumstances, e.g. if organic matter such as coal grains is present in the clay. If wells drilled or dug are closed airtight, in absence of oxgyen, the organic material causes reduction of sulphates and the SO_3-content of the water decreases substantially within a few days.

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(1) A. VENDL : Uber das Grundwasser des Làgymànyos (Subsoil Water of Làgymànyos) Hidrolôgiai Kôzlôny (Periodical of Hydrology) X. Budapest. 1930, pp. 3-17.

In November 1947 Prof. *Jáky* had this site examined by 6-15 m deep borings in order to ascertain the carrying of the subsoil for foundation works. Some water samples showed SO_3-contents of 459-4930 mg/l. In some holes ground water level and the elevation of the blue, unaltered Kiscell Clay ware also determined. The former showed differences of 0,95 m and the latter was 2.46–3.58 referred to the 0-point of the Danube.

The number of wells was variable; exhausted wells, which no more yielded water of sufficient concentration, were filled up. At present there are 15 wells : the water level therein was on June 3rd, at medium Danube level, from 2,51 to 5 m.

From a great number of analyses the following are selected to account for the chemical composition of the mineral water of these wells :

Table 1. — Name of Well

1000 g water contains	Aesculap	Màtyàs III	Heinrich	Deàk F.	Erzsébet Bath	Test	New
						Well	
g							
K	0,0048	0,3511	0,0972	0,1276	0,0860	0,0579	0,0740
Na	6,0804	5,3993	6,7908	6,3736	7,5773	7,9912	5,0844
Mg	3,4561	3,0033	2,7228	3,5983	3,5214	3,8514	2,2897
Ca	0,6217	0,4136	0,3155	0,4452	0,4916	0,6969	0,3709
Fe	0,0047	trace	0,0097	0,0062	0,0056	—	0,0040
Al	0,0186	»	0,0015	—	—	—	—
SO_4	24,6930	22,3805	20,8548	25,2208	27,8990	29,2893	18,7312
Cl	1,7620	1,3119	1,3458	1,7495	1,7069	1,7509	0,9078
HCO_3	0,8981–CO_3	0,6846	2,0410	/$CO_3$0,9980	1,0468	2,5341	0,8613
SiO_2	0,0097(SiO_3)	0,0109	0,0213	0,0063	0,0160	—	—
H_2SiO_3	—	—	—	—	—	0,0100	0,0142
Total :	37,5714*	33,5552	34,2004	38,8727***	42,3506	46,1817	28,3375
Free CO_2		0,0335	0,3753**				1,02641
d =	1,0354	1,03295	1,03074				
Analyzed by :	J. Molnàr	L. Ilosvay	J. Molnàr	K. Than		K. Emszt	I.Finàly
Éear :	1878	1897	1872	1891		1929	1935

* Including Mn = 0,0207 + H_4N = 0,0016.

** Free and half fixed CO°.

*** Including : 1/2 CO° = 0,3472.

The sulphate content varies with the water level of the Danube and with conditions of evaporation. SO_3 in mg/l was determined from well bottom samples taken on June 2nd, 1947 as follows (Analyst : A. Almàsy). Well No 8 : 6810 mg; No 9 : 10380 mg; No 10 : 14860 mg (on May 10th this content was 11600 mg); No 13 : 2015 mg; No 14 : 4300 mg;No 16 : 6210 mg;No 17 : 2890 mg;No 18 : 6400 mg; No 19 : 15670 mg; No 20 : 15770 mg; No 21 : 12830 mg; No 22 : 14100 mg.

Well mouth elevation above sea level is from 101.77-104.36 m; wells are on the average 50-60 m distant from one another, but this distance is sometimes less and sometimes more, up to 300 m. Wells are operated on the average for 15-20 years but those in strong demand are generally exhausted in 10 years, because their sulphate content sinks to a degree which renders them unsuitable as medicinal water. The total quantity of pyrite has been oxydized around such wells, so that the clay supplies water of insufficient concentration, but for the case when water arrives to the well from a distance with sufficient concentration already. Exhausted wells are filled up.

The number of wells on the Örsöd « Hunyadi Jànos » estate has also been variable. In 1881 there were 40 wells, in 1885; 70 wells of which 60 were in operation. In 1929 there were 166 wells, in 1938 : 179 wells there, partly out of operation. The wells are on the average 6-8 m, old wells often only 5 m deep. The average section of wells is shown on fig. 4. The bitter mineral water soaks from the yellow clay chiefly at its contact with the deeper blue clay into the well. The clay on the bottom of the wells is very hard to dig, it is comparatively dry and contains but little sulphate. The wells have 1,9-2,5 m diameter and are lined with Tétény Sarmatian limestone pervious to water.

Part of the ground water percolates through pores and cracks of the bog-earth covering the clay, to the surface. In dry weather efflorescence of Glauber's and Epsom salts are sometimes visible on the soil.

The wells of Örmező also reach somewhat deeper than the contact of the yellow and of the unaltered blue clay. Their cross section and dimensions are essentially the same as those of Orsöd wells.

Of very numerous analyses the following examples account for the chemical composition of these mineral waters.

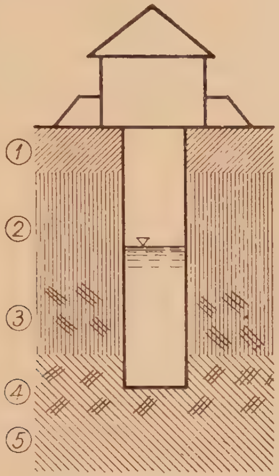

Fig. 4. — *Section of Örsöd bitter water wells.*
1 = Clay transformed into humus; 2 = yellow clay; 3 = yellow clay with blue spots; 4 = blue clay with Yellow spots; 5 = blue clay.

«Hunyadi János» Bitter Mineral Water

1,000 g water contains g				
K_2SO_4	0,1206	0,0849	0,132943	0,0174
Na_2SO_4	22,5514	15,9148	19,662123	23,1116
$MgSO_4$	22,3500	16,0158	18,449451	17,7485
$CaSO_4$	—	—	1,321938	0,1582
$SrSO_4$	—	—	—	0,0090
Na Cl	1,7048	1,3050	1,424068	1,8452
$NaHCO_3$	0,6760	—	—	—
$Sr(HCO_3)_2$	0,0270	—	—	—
$Ca(HCO_3)_2$	0,7967	0,9330	—	1,4495
$Fe(HCO_3)_2$	0,0006	—	—	trace
Na_2CO_3	—	0,7960	—	—
$MgCO_3$	—	—	0,731347	—
$FeCO_3$	—	—	0,002059	—
$Al_2O_3+Fe_2O_3$	—	0,0042	—	—
SiO_2	0,0106	0,0011	0,011212	0,0127
Total :	48,2377	35,0548	41,735147	44,3521
Free and half fixed CO_2	5,5225	0,5226	217.44 cm^3	
d =		1,03323		
Analysed by :	R. Bunsen	J. Liebig	R. Fresenius	E. Ludwig
in the year :	1876	1870	1878	1885

«Feremcz József» Bitter ...Mineral Water

K	0,08896
Na	4,0671
Li	0,00004
Ca	0,4140
Sr	0,00008
Mg	3,4321
Fe	0,000044
Mn	0,00002
Al	0,0048
SO_4	21,3128
Cl	0,6274
Br	0,00014
HCO_3	1,0755
NO_3	0,0042
HPO_4	0,00055
SiO_3	0,01435
Total :	31,04208
Meta-titanic Acid	0,00001
CO_2 dissolved	0,0154
Analysed by :	J. Jendrassik
in the year :	1932

Since 1946 wells of Orsöd and Örmezö are controlled by the *Section of Well Exploration* under the Author's control. The result of observations has hitherto been as follows :

1) At high groundwater level the water level in the wells rises as well.
2) With the rise of water level the concentration of bitter mineral water decreases substantially.
3) The concentration of the water rises usually only when the Danube water level has been low during a longer period.
4) In the wells the upper section of the water column has maximum, the bottom section minimum density.
5) The temperature of water is different at different depths. In the cold months, the bottom of the water column, in warmer months, the top of the water is warmest.

Bitter mineral water of the Erzsébet Brine Bath wells is used for medicinal baths, water of the Örsöd and Örmezö region is marketed under the names of «*Hunyadi János*», «*Ferencz Jozsef*» and «*Apenta*» Bitter Mineral Water as purgative mineral water.

11) THE THERMAL SPRINGS OF NEW ZEALAND

by

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SUMMARY.

Hot springs are scattered through the central and northern part of the North Island of New Zealand, with two isolated localities in the east. Those in the South Island are limited to a strip in the north central portion, and one isolated spring on the east coast. Altogether 57 localities have been listed, each including springs from one to several hundreds in number.

In all cases the thermal waters are regarded as being largely, and in some cases entirely, of meteoric origin, and in practically all cases the springs lie on the lines of faults, and the waters rise through the associated fracture zones. The mechanism of flow is controlled by two agencies — the lower density of hot water relative to the surrounding cold water, and the fairly general presence of gas, which also decreases the density of the water.

The greatest number of hot springs, and also the most active, lie along a narrow strip in the North Island, extending from the National Park volcanic mountains north-east to White Island, an active volcano 28 miles from the coast. There have been historic and recent volcanic eruptions along this zone, and there can be little doubt that the heat source for these springs is magma at no great depth beneath the surface. These springs are believed to contain water and chemical constituents of magmatic origin, but no attempts at calculating the amount have been made.

The discharge of gas from springs in this central region is generally greater than elsewhere, as is also the variety in the types of springs. Where ground water level is at or near the surface and the permeability of the rocks is high, alkaline springs usually exist. Where water level is entrenched, and the permeability near the surface is low, the movement of the ground water is considerably restricted, and acid conditions are usual, especially where gas discharge is abundant. Hydrogen sulphide gas is a common constituent of the spring gases, and oxidation of this where ground water movement is retarded results in the formation of sulphuric acid which attacks the surface rocks to form sulphates. These conditions are believed to be superficial, and alkaline conditions are believed to exist at lower levels, except in cases such as White Island, where the magma is sufficiently close to the surface for the direct condensation of acid emanations.

The remaining hot springs in the North Island to the north-west of the volcanic group lie close to volcanic vents active during the Pliocene and Pleistocene. The heat source for the thermal springs in these cases is likely ultimately to be magmatic. The Waiwera and Helensville springs are not close to volcanic vents but are believed to lie on faults, though it appears likely that they have the same heat origin as the others.

The two hot springs on the east coast of the North Island are associated with structural « highs » and are structurally controlled. The South Island springs with one exception are associated with faults that dislocate high mountain ranges of Mesozoic and Palaeozoic rocks, in a region where uplifts of 10,000 feet and more have occurred about the end of the Pliocene or in the early Pleistocene. The springs emerge in the valleys of rivers that have cut deeply into these elevated blocks. The heat is thought to be from two sources. It is likely that the rise of the geotherms that accompanied the uparching is still operative, and large bodies of rock adjacent to the fault planes were heated by friction during the faulting movements.

One hot spring at Lyttelton lies within an erosion caldera that forms the harbour. Intermittent volcanic activity has continued here from Cretaceous till late Tertiary times, so that a magmatic origin for the heat in this case is postulated.

The report is accompanied by 52 chemical analyses of typical waters from most of the thermal springs localities, and two maps show the areal distribution of the springs.

Innot Hot Springs.

This is situated in the bed of Nettle Creek, 8 ½ miles in a direct line E. by N. of Mount Garnet.

The surrounding country is composed of slates, schists, and quartzites, intruded by granite; both metamorphics and granite being intersected by dykes of felsite.

The spring itself issues through the sands and gravels of the creek bed. Gas, with a slightly sulphurous odour, rises through the water.

No deposits exist around the mouth of the spring, but an exceedingly thin white coating has been noticed on many of the pebbles near the overflow.

In 1900, B. Dunstan recorded old deposits of siliceous sinter close to the Spring, covered by recent alluvium. Samples of this sinter showed incrustations of silica around recent vegetable remains. The temperature of the issuing water was given in 1891 as 189° F.; in 1899 the normal temperature was stated to be 158° F., with variations up to 168° F. Two analyses of the water are available.

	1896 grains per gallon	*1899 grains per gallon*
Total Solids	46.80	41.6
Iron and Alumina	0.40	0.8
Calcium carbonate	1.86	2.6
Magnesium carbonate	1.70	—
Sodium carbonate	15.34	8.6
Sodium chloride	19.60	19.1
Sodium sulphate	—	3.3
Silica	7.90	7.2
Ammonia	0.16	—
Lithia	trace	—

A sample of the gas issuing in 1940 contained CO_2, 1.8%; Inert gas 98.2%,

Einasleigh (Talaroo)

The springs are near the Einasleigh River, about 30 miles in a direct line N.N.W. from Einasleigh township.

They occur 1/4 mile from the river in a mound which was stated by R.I. Jack in 1892 to be a dome-shaped mass, 15 feet in height and 200 yards in circumference, rising out of a plain of shallow recent alluvial deposits, beneath which, in the banks of the river, are exposed gneisses and schists forming part of a series of ancient highly contorted rocks occurring in the surrounding country. R.L. Jack stated that there were several orifices within the mound. Bubbles of gas rise continually, and in the spring at the apex of the mound he noticed a rush of bubbles at intervals of 2 1/2 minutes.

The mound, which is a dirty white or drab colour, is composed of carbonate of lime, and is covered by semi-circular or oval basins forming a series of steps from the summit to the edge of the dome. The overflowing waters pours into these cups one after another.

The calcareous sinter is light and spongy at the surface, but is tolerably compact in the interior. Loose in the bottoms of the cups are needles of calcite.

Jack recorded that the gas given off had a distinct odour of hydrogen sulphide. On his visit in 1889 Jack referred to the water as having a very disagreable taste, but on a later visit in 1898 he found the water almost tasteless.

The temperature of the issuing water was given in 1911 as 48° F., and in 1929 the maximum temperature was found to be 145.5° F.

Three analyses of samples of water have been made.

(a) *1892*

Carbonates of Calcium and Magnesium	6.25 grns per gln.
Carbonates of Sodium and Potassium	15.94 grns per gln.
Chlorides of Sodium and Potassium	32.61 grns per gln.
Total fixed salts	54.80 grns per gln.
Volatile matter	2.80 grns per gln.
Total solids	57.60 grns per gln.

Sulphuric Acid — trace
Hydrogen sulphide — 2.19 grns per gallon.

(b) *1929*

Total solids	56 grns per gln.
Chloride (as sodium chloride)	29 grns per gln.
Alkalinity (as sodium carbonate)	12 grns per gln.
Sulphates (as sodium sulphate)	9 grns per gln.
Hardness (degrees Clark)	3 grns per gln.

(c) *1947*

Total solids	50.0 grns per gln.
Calcium sulphate	6.5 grns per gln.
Magnesium sulphate	0.3 grns per gln.
Sodium sulphate	3.0 grns per gln.
Sodium carbonate	11.2 grns per gln.
Sodium chloride	28.4 grns per gln.
Hardness	5.0 grns per gln.

A sample of the calcareous sinter gave on analyses

Moisture	0.3 %
Loss on ignition	43.6 %
Insolubles	1.3 %
Iron oxide and alumina	0.6 %
Calcium oxide	53.5 %
Magnesium oxide	0.3 %
Organic matter present	
Chlorides present	
Sulphates present	

A sample of the gas collected in 1929 contained

CO_2 — 0.3 %
Inert gas — 99.7 %

Saxby River, a tributary of the Flinders, 80 miles in a direct line N.E. of Cloncurry.

Hot springs were recorded many years ago on the southern side of Mount Brown, near the Saxby River. The water, with a temperature of 120° F., was stated to be standing in a large basin on a mound raised many feet above the surrounding plain. The area is one of Cretaceous marine sediments, but granite outcrops near the Flinders River further north and may occur also on Mount Brown.

A white incrustation on the overflow from the spring gave the following analysis

Water	27.793 %
Silica	0.600 %
Chlorine	3.369 %
Sodium	2.183 %
Carbonic acid	33.735 %
Soda (sodium carbonate)	31.690 %
Sulphuric acid	present

No change in level or temperature was observed between 1865 and 1884.

Musgrave Telegraph Station. 125 miles W.N.W. of Cooktown.

A small hot spring, with temperature 100.5° F. was reported in 1901, from which considerable quantities of hydrogen sulphide were evolved. The spring is in granite country. Calcereous sinter was found covering an area 300 yards by 50 yards in the vicinity of the spring, though none was then being deposited.

Inniskillen (between Tambo and Blackall)

Small hot springs were reported many years ago on the Barcoo River, below Inniskillen, but no particulars are known.

Thargomindah

A hot spring is recorded in this district on the margin of a ridge of granite and gneiss forming an island in the Cretaceous sediments of the Artesian Basin. No details are known as to exact locality or extent of the occurrence.

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13)

THE THERMAL WATER OF THE UNION OF SOUTH AFRICA AND SOUTH WEST AFRICA

by

LESLIE E. KENT

(Communicated by permission of the Honourable the Minister of Mines)

In the Union of South Africa seventy-three springs are regarded as thermal, i.e., issuing at temperatures exceeding 25°C. the year round. Those above 37°C. are classed as hot, whilst a special group styled scalding waters has been created for those with temperatures exceeding 50°C. The combined discharge is estimated to be 8,000,000 gallons (36,290 cbm.) a day, one spring alone being responsible for almost a third of this total. Thermal water also flows from several boreholes, in five of which it was struck at depths greater than 2,500 feet (762 m.).

Twenty-five analyses representative of all the types of thermal water known in the country are given, and of these seventeen have not been published previously. The waters contain up to 86 mgm. per litre of silica, relatively little of which is believed to be present in either the ionic or the colloidal condition, a view contrary to that generally held.

Sulphuretted waters, which are defined as containing more than 10 mmg. of dissolved hydrogen sulphide per litre, are uncommon in South Africa. It is suggested that the terms « moderately sulphuretted » and « slightly sulphuretted » should be used for those containing from 5 to 10 and from 1 to 5 mgm. per litre of the dissolved gas respectively.

Fluorine is a prominent minor constituent, up to 13.2 mgm. per litre being present. Strontium, an element seldom sought in waters, occurs to the extent of 7.2 mgm. per litre in water from a deep borehole near Trompsburg.

Saline residues from the waters were spectrographically examined by Dr. B. Wasserstein of the Geological Survey. Selected minor elements seemed to characterise waters issuing from geological formations. Rough quantitative data indicated that the elements Li, Sr and Ba did not form more than 0.1 per cent. of the residues, while Pb and Mo both had maximum values around 0.005 per cent.

The gas associated with the waters is of two types; one is mainly air from which most of the oxygen has been abstracted and the other mainly methane, accompanied in some cases by hydrogen. Thirteen new analyses are given.

Spring deposits are not common and are represented only by ferruginous ochres and sinters formed by springs in the Cape system. Manganese is an important constituent of these deposits.

In most cases a close relationship in composition exists between the thermal waters and water struck in shallow boreholes or issuing as cold springs. When differences other than in degree of concentration were noted, it was found that the thermal waters were characteristic of the deeper underlying rock formations.

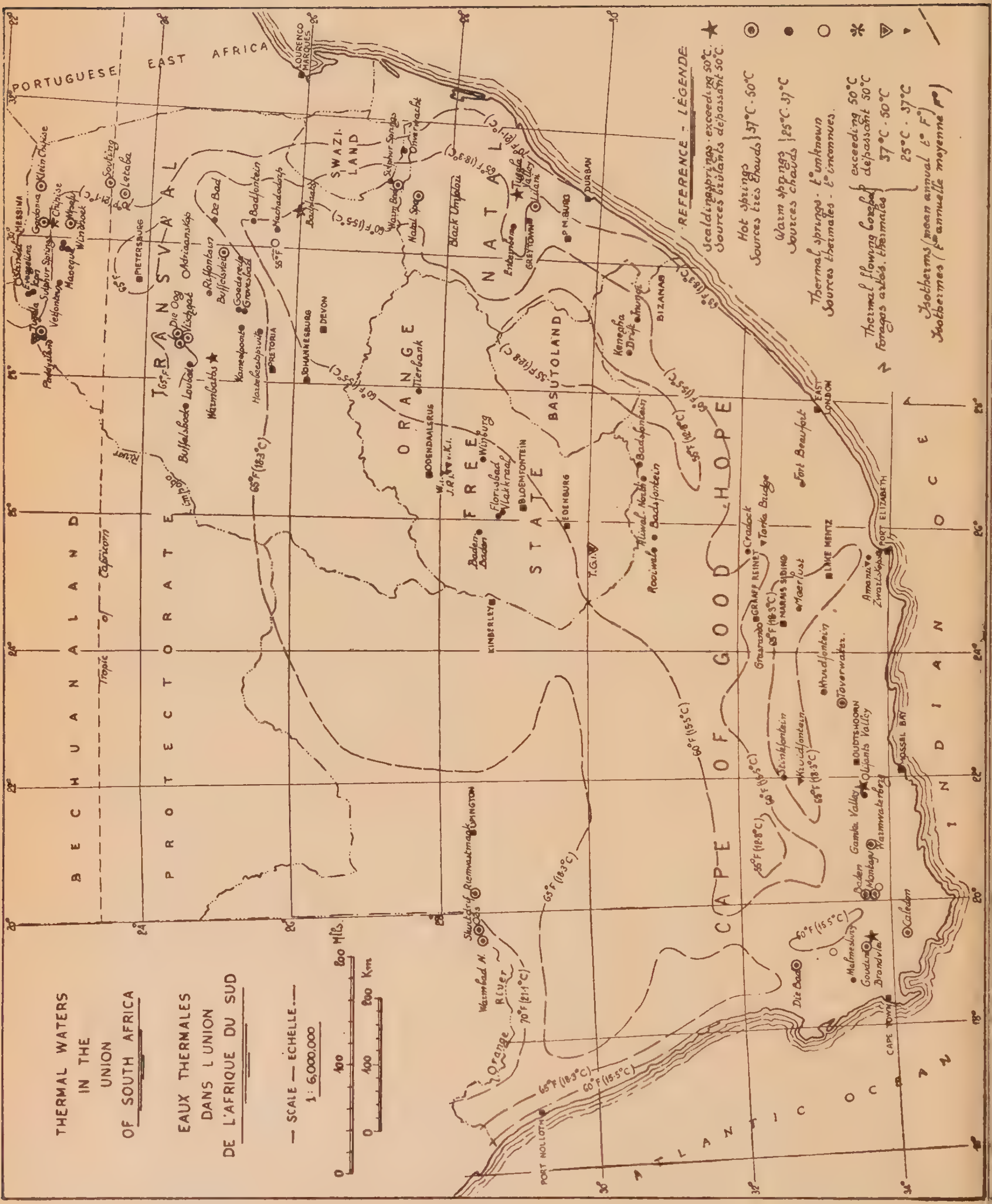
All the thermal waters are considered meteoric. Geological sections illustrate the structures upon which the emergence of thermal waters depends. These are artesian basin, artesian slope, barrier, and faults. The source of the heat is probably the normal geothermal gradient in the crust. It is calculated that the total amount of heat being brought to the surface by the South African thermal springs is approximately 5.8×10^{-10} cal.cm.^{2}sec. or about a two-thousandth of that normally dissipated by conduction through the crust.

Less is known about the thermal springs of the Territory of South West Africa Twenty-four springs are known to be thermal, of which nine are scalding, a much higher proportion than in the Union of South Africa. Warm water has been struck in many fairly shallow boreholes.

A few analyses are given, all old. The potassium content greatly exceeds that of the thermal waters of the Union; this is attributed to the higher potassium content of the granites of South West Africa.

The Windhoek springs have deposited extensive layers of calcareous tufa.

The South West African thermal springs, too, are believed to represent the return of meteoric water to the surface mainly through fault fissures and in some cases fractures associated with volcanicity and possibly of Cretaceous age.



THERMAL WATERS IN THE UNION OF SOUTH AFRICA

EAUX THERMALES DANS L'UNION DE L'AFRIQUE DU SUD

— SCALE — ECHELLE —
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PART I.—THE UNION OF SOUTH AFRICA

I. Introduction

For a non-volcanic country the Union of South Africa is relatively well-endowed with thermal springs. There are no definite thermal regions, and with few exceptions the springs occur sporadically throughout those parts of the country where the annual rainfall exceeds 10 inches (25.4 cm.), i.e. in all save the arid North of the province of the Cape of Good Hope.

The natural hot waters of the western Cape Province attracted the early explorers and travellers, and Caledon and some of the other springs have been health resorts since the early 19th century. The first scientific account of them, by T. E. Gumprecht,[15] is now almost a hundred years old. His descriptions included temperature measurements but no chemical analyses. The first attempt at a chemical classification of all the known South African « medicinal » springs was published by M. M. Rindl[26] in 1916, and the work was continued as a series af papers. These deal essentially with the chemical composition of the waters. In most geological accounts of areas where thermal springs are found they have either been completely ignored, or their existence has merely been noted. Reasonably detailed descriptions by geologists of only the Liliani,[7] Amanzi,[8] Tugela valley,[14] Letaba,[17] Loubad,[18] Souting,[19] and Buffelshoek[20] springs are available in print, but several others have been investigated recently, and it is hoped within a few years to publish a Geological Survey Memoir on the thermal waters of South Africa.

Artesian supplies of thermal water have been struck in several boreholes : most of them are deep, having been drilled in search of gold reefs or oil. Warm waters have also been encountered in a few of the deeper mines.

II. CLASSIFICATION OF THERMAL WATERS ON THE BASIS OF TEMPERATURE

It appears to be largely a matter of opinion where the dividing line should be drawn between thermal and non-thermal waters. Kirk Bryan[3] mentioned that it is usually taken as 70°F. (21.1°C.), but that 20° to 25°F. (11.1°-13.9°C.) above the mean for the region might be preferable. In the most recent general account of the thermal springs of the United States, however, some are included « whose temperatures may not be more than 10°F. (5.6°C.) above the mean annual temperature of the locality where they issue."[31] In an official Canadian publication[10] thermal springs are defined as those issuing at above 70°F. (21°C.)the year round, those above 90°F. (32.2°C.) being described as hot. In Southern Rhodesia, the Union's neighbour to the north, the dividing temperature of 20°C. (68°F.) has been chosen.[22] This follows climatological usage as 68°F. is regarded as the temperature separating « mild » from « warm », whereas 80°F. (26.7°C.) is used to separate « warm » from « hot ». Most balneotherapists regard 80°F. as the division between cool and tepid waters, and they class waters between 92° and 98°F. (33.3°-36.7°C.) as warm, between 98° and 104°F. (36.7°-40.0°C.) as hot, and above 104°F. as very hot. On an Italian classification[23] waters below 20°C. are regarded as cold, those between 20° and 30°C. as hypothermal, those between 30° and 40°C. as thermal, and those above 40°C. as hyperthermal. The scheme proposed by the International Society of Medical Hydrology is rather similar—waters below 20°C. are termed cold, between 20° and 37°C. subthermal, from 37° to 42°C. thermal, and above 42°C. hyperthermal.

The mean annual air temperatur of the Union of South Africa ranges between approximately 55° and 70°F. (12.8-21.1°C.). Most of the country falls between the 55° and 65°F. isotherms, but many of the thermal springs lie in the 65° to 70°F. (18.3°-21.1°C.) belt[24]. Thus it would appear to be inadvisable to set a limit below 25°C. (77°F.) as the division between thermal and non-thermal waters. Even for geologists human blood-heat appears a most satisfactory temperature for dividing warm from hot waters. In a subtropical country like South Africa warm waters therefore have a range of 12°C. A new dividing line is proposed at 50°C. (122°.) to separate hot from scalding waters. On this basis the hot waters have a range of 13°C. Thus the classification proposed is :

Warm	25°—37°C.
Hot or hyperthermal	37°—50°C.
Scalding	above 50°C.

Natural waters with temperatures between the mean air temperature and 25°C. may be classed as tepid. Provision may be made in the classification for boiling and superheated waters, but these are not known in South Africa.

The Union of South Africa has an area of 472,550 sq. miles (122,394,230 hectares). On the basis of the above classification there are seventy-three known thermal springs and nine thermal artesian boreholes divided as follows :—

Springs.		Artesian Thermal Boreholes.
Warm	38	6
Hot	24	2
Scalding ..	6	1
Temperature unknown .	5	
	73	9

III. DISCHARGES OF SPRINGS

The spring at Brandvlei in the south-west of the Cape Province is by far the strongest thermal spring in South Africa, having a daily discharge of 2,430,000 gallons. (11,022 cbm.). Of the others only Aliwal North exceeds a cusec (540,000

SOUTH AFRICAN THERMAL SPRINGS

	Spring	Latitude °S.	Longitude °E.	Tempe-perature °C.	Flow Imp. Galls. 24 hrs.	Flow Cu. metres–24 hrs.	Chemical Classification (Bond System).*	Geological structure determining origin.	Observer of temperature and flow, or (reference) and date.
SCALDING SPRINGS. SOURCES BRULANTES.	Brandvlei . . .	33° 43′	19° 25′	64.2	2,430,000	11,022	E	Middle Cretaceous fault in Cape system (Devonian-Lower Carboniferous).	C. Preston, 1947 (flow) L. E. Kent, 1947 (temp.).
	Chipise	22° 37′	30° 10′	57.2	100,000	454	D, B	Intersection of two post-Permian faults in Upper Karroo.	H. N. Visser, 1948.
	Tugela valley	28° 52′	31° 0′	52—53	—	—	A (SO_4)	Fault in Archaean amphibolites and gneisses.	T.W. Gevers, 1942.
	Warmbaths	24° 53′	28° 18′	51.9	160,000	726	D	Post Karroo fault in Bushveld granite (pre-Cambrian).	(M. Rindl, 1931.)
	Olifants valley	33° 40′	21° 46′	50—51	144,500	655	E (Fe)	Middle Cretaceous fault in Table Mountain series, Cape system.	L.E. Kent and P.G. du Plessis, 1947.
	Badplaats . .	25° 57′	30° 35′	50	180,000	817	D	Fracture in Archaean granite.	Baths Superintendent, 1948.
					3,014,500	13,674			
HOT SPRINGS SOURCES TRES CHAUDES.	Caledon	34° 15′	19° 27′	49	180,000	817	E (Fe)	Middle Cretaceous fault in Cape system.	(Rindl, 1931.)
	Warmwaterberg	33° 44′	20° 54′	45.6	174,000	789	E (Fe)	Fault at nose of anticline of Table Mountain series.	L. E. Kent, 1947.
	Warmbad Noord	28° 32′	19° 33′	44.4	15,000	68	—	Fracture zone in Archaean granodiorite.	D. J. L. Visser, 1943.
	Natal Spa . .	27° 32′	30° 52′	44.4	64,800	294	D	—	Owners, 1936.

(*) Bond classes the underground waters of South Africa into five groups designated A to E. For the significance of these letters, see page 209 of this paper. The small symbols added in brackets, i.e., SO_4, Fe and S, denote respectively sulphate-rich waters, chalybeatic waters, and waters containing more than 1 mgm, per litre of H_2S.

HOT SPRINGS.
SOURCES TRES CHAUDES.

Spring	Latitude °S.	Longitude °E.	Temperature °C.	Flow. Imp. Galls.– 24 hrs.	Flow. Cu. metres– 24 hrs.	Chemical Classification (Bond System).*	Geogical structure determining origin.	Observer of temperature and flow or (reference) and date.
Toverwater..	33° 24′	23° 10′	44.3	216,000	980	E (Fc)	Middle Cretaceous fault in Cape system.	L.E. Kent(temp.) and D. Wallace (flow), 1946.
Welgevonden	24° 27′	28° 34′	44	12,500	57	—	Diabase dyke along post-Karroo (post-Permian) fault in Rooiberg felsites.	J. W. von Backström, 1938.
Souting	23° 25′	30° 51′	43.9	30.000	136	A	Fault in Archaean granite	T.W. Gevers,1945
Mpefu	22° 55′	30° 10′	42.8, 43.7	—	—	E	Pre-Karroo (pre-Carboniferous) fault.	L.E. Kent, 1939.
Die Bad.....	32° 45′	19° 2′	42.2-43.2	—	—	—	Barrier spring on west limb of syncline in Cape system.	M. Rindl, 1916.
Tugela †	22° 34′	28° 37′	42.8	2,650	12	A	Joints in Archaean gneiss	L.E. Kent, 1948.
Letaba......	23° 39′	30° 40′	40.4-42	91,000	413	A	Dolerite dyke in Archaean gneissic granite.	L.E. Kent, 1942.
Black Umfolosi	28° 2′	31° 18′	41	—	—	B (S)	Fissures in Dwyka tillite (U. Carboniferous) overlying Archaean rocks.	G. Molengraaff, 1898.
Goudini	33° 40′	19° 16′	40.1	—	—	E	Middle Cretaceous graben in Cape system (see fig. 2)	L.E. Kent, 1945.
Lilani	29° 7′	30° 51′	38—40	—	—	D (S)	Fracture zone in Archaean gneisses and amphibolites	A.L. du Toit,1916.
Die Oog	24° 26′	28° 37′	39.5	32,000	145	E	Diabase dyke along post-Karroo fault in Rooiberg felsites.	J.W. von Backström, 1938.
Vischgat ..	24° 34′	28° 36′	39.5	20,000	91	—	Post-Karroo fault in Bushveld granite.	» »
Skuitdrif Oos	28° 30′	19° 43′	±38	—	—	—	Fracture zone in Archaean granite.	D.J.L. Visser, 1943.
Gordonia..	22° 35′	30° 10′	37.7	—	—	—	Fault in Archaean gneiss.	H.N. Visser,1948

† Borehole near spring flows at the rate of 15,600 galls. (71 cbm.) per 24 hours, the temperature of the water being 48.9°C.

Spring.	Latitude °S.	Longitude °E.	Temperature °C.	Flow. Imp. Galls. 24 hrs.	Flow. Cu. metres– 24 hrs.	Chemical Classification (Bond System.)*	Geological structure determining orifgin.	Observer of temperature and flow, or (reference) and date.
Aliwal North	30° 43′	26° 43′	36.9	840,060	3,813	A	Dolerite dyke in Beaufort series (L. Triassic).	S.V.R. Lewis, 1927.
Loubad	24° 36′	28° 11′	27.1-34.1	414,300	1,879	E	Artesian basin in Water–berg system (pre-Cambrian ?) (See Fig. 2).	L.E. Kent, 1940-1943.
Gamka valley	33° 40′	21° 43′	32.3-33.2	65,600	297	E(Fe)	Middle Cretaceous fault in Table Mountain series.	L.E. Kent and P .G. du Plessis, 1947.
Malmesbury	33° 28′	18° 43′	32.9	180,000	816	A (S)	Fracture zone in Cape granite (pre-Cambrian ?)	Municipality.
Grovesbad .	25° 21′	29° 2′	32.8	—	—	—	Fissures in Bushveld granite.	L.E. Kent, 1942
Evangelina .	22° 25′	29° 11′	32.5	—	—	A	Diabase dyke in Archaean gneiss.	L.E. Kent and O.R. van Eeden, 1941.
Sulphur springs	27° 11′	31° 6′	31	120,000	544	E (S)	—	J.K. Marais, 1946
Cradock	32° 8′	25° 27′	29–31.3	18,300	83	D (S)	Dolerite dyke in Beaufort series sediments.	L.E. Kent, 1946
Buffelshoek . .	24° 34′	27° 36′	30.6	17,000	77	D	Diabase dyke as barrier on artesian slope of Bushveld granite. (See Fig. 2).	L.E. Kent and H.D. Russell, 1944.
Rooiwal ...	30° 52′	25° 35′	30	56,000	254	D	Dolerite dyke in Beaufort series.	D.J.L. Visser, 1947.
Badsfontein .	30° 53′	25° 47′	25.5-30.0	51,000	231	B	Dolerite dyke in Beaufort series.	D.J.L. Visser, 1947.
Winburg ...	28° 33′	26° 55′	29.5	4,000	18	A	—	M. Rindl, 1918.
Vetfontein . .	22° 47′	29° 23′	29.5	—	—	—	Post-Karroo fault.	L.E. Kent, 1945.
Knegha Drift	30° 28′	28° 39′	29.3	—	—	—	Dolerite dyke in Beaufort series.	L.E. Kent, 1947
Riffontein . .	24° 50′	29° 18′	28.3-29.0	8,000	36	A-B	Fault in Rooiberg series pre-Cambrian).	L.E. Kent, 1942, and J. T. Wessels 1948.
Machadodorp	25° 39′	30° 15′	27.5-28.5	45,000	204	D	—	(M. Rindl, 1916.)
Fort Beaufort	32° 50′	26° 40′	27.0-20.9	17,300	79	D (S)	Fissure in Beaufort series	L.E. Kent, 1947.
Stinkfontein	32° 40′	21° 59′	28.7	7,500	34	D (S)	Major joint in syncline of L. Beaufort sediments).	L. E. Kent, 1949.
Etembeni ...	28° 51′	30° 29′	28	—	—		Dolerite sill in Ecca series (Permian) shales.	A.L. du Toit, 1931
Grasrand....	32° 19′	24° 27′	26	3,600	16		Joints in L. Beaufort sediments.	L.E. Lent, 1949.

gallons or 2,450 cbm.) per day. Brandvlei is also the hottest spring in the country, while Aliwal North has a higher temperature than the other springs in the north-eastern Cape Province and adjacent Orange Free State.

Of the springs that have been gauged only the two mentioned above attain third magnitude on Meinzer's classification, whereas there are eight of fourth magnitude, and fourteen of fifth magnitude.

A preliminary estimate gives the combined discharge of all the thermal springs in the Union of South Africa as 8,000,000 gallons (36,290 cbm.) per day. There are few figures available from other countries that can be quoted for comparison. In the United States of America, which has an area six and a half times greater than that of the Union, the thermal springs have a daily discharge of 720,000,000 U.S. gallons. (2,719,000 cbm.).[31]

In the Table (pp. 205, 206, 207) the results are given of recent measurements of flow and temperature, together with geological and chemical data. Only those springs for which reliable information is available are included.

The flows of the thermal artesian boreholes are given in the Table on page 217.

IV. CLASSIFICATION OF THERMAL WATERS ACCORDING TO CHEMICAL COMPOSITION

The therapeutic value of thermal waters is often ascribed to constituents of relatively minor importance chemically, and waters are commonly classed as « arsenical, » « lithia, » « iodine, » etc. Under the heading « sulphur springs » Rindle[26] included several containing less than 1 mgm. per litre of dissolved hydrogen sulphid and hydrosulphide ions. According to Matsuo[21] sulphur springs should contain at least 10 mgm. per litre of free hydrogen sulphide. It is not known if this definition has been found generally acceptable; if it has, then as far as our present knowledge goes there is only one thermal sulphur spring in South Africa, Stinkfontein in the Cape Province. In special samples taken at various springs smelling quite strongly of hydrogen sulphide this fugitive constituent was fixed immediately as cadmium sulphide. Despite this precaution the highest figures obtained for dissolved hydrogen sulphide were 16.0 and 6.2 mgm. per litre—for the springs at Stinkfontein and near Fort Beaufort respectively. At Badplaats and Zwartkops, where the odour of hydrogen sulphide is only very slight, analyses showed only 0.2 and 0.5 mgm. per litre of the dissolved gas respectively.

In view of the importance attached by balneotherapists to the presence of dissolved hydrogen sulphide in waters, the following definitions are suggested :—

Sulphuretted waters contain over 10 mgm. per litre dissolved H_2S (including HS')
Moderately sulphuretted waters contain between 5 and 10 mgm. per litre dissolved H_2S (including HS')
Slightly sulphuretted waters contain between 1 and 5 mgm. per litre dissolved H_2S (including HS')

Sulphurous waters, i.e. those containing dissolved sulphur dioxide, are not known from South Africa, but the same limiting values could be used.

Waters that deposit hydroxides of iron present rather a difficult problem as regards classification. The $Fe^{\cdot\cdot}$ content of these so-called chalybeate springs is often surprisingly low. A recent series of analyses gave values of 0.7, 2.3, 2.9, 3.8, and 7.7 mgm. per litre for the $Fe^{\cdot\cdot}$ content of the Warmwaterberg, Olifants valley, Caledon, Zwartkops, and Gamka valley springs respectively. Manganese is actually present in slight excess over iron in the Warmwaterberg, Caledon and Olifants valley thermal waters.

The iron is almost certainly present as ferrous bicarbonate, and, as is well known such waters seldom contain more than 30 mgm. per litre of $Fe^{\cdot\cdot}$. Sulphate waters, on the other hand, may contain as much as 1 gm. per litre of $Fe^{\cdot\cdot}$ in the form of ferrous sulphate. The only thermal springs of this latter type known in the Union are at Amanzi[8] (Balmoral), near Port Elisabeth, and Inungi; their waters contain 48.5 and 52.9 mgm. per litre of $Fe^{\cdot\cdot}$ respectively.[16]

With the exception of these two it is proposed to term the iron-depositing springs « chalybeatic » rather than « chalybeate».

In his « Geochemical Survey of the Underground Water Supplies of the Union of South Africa, « published in 1946, Bond[2] recognised five distinct classes :

A — Highly mineralised Chloride-Sulphate waters.
Total solids$>$1,000 mgm. per litre. Cl$'>$27 per cent. $SO_4''>$5 per cent.
B — Slightly saline Chloride Waters.
Total solids$>$300$<$500 mgm. per litre. Cl$'>$27 per cent. $SO_4''<$3 per cent.
C — Temporary hard Carbonate Waters.
Total solids$<$800 mgm. per litre. pH$>$7.6.
D — Alkaline Soda Carbonate Waters.
Total solids$<$1,000 mgm. per litre.
Na_2CO_3 or $NaHCO_3>$15 per cent. Permanent hardness, nil.
E — « Pure » Waters.
Total solids$<$150 mgm. per litre. pH$<$7.1.

Bond's[2] investigations were mainly in connection with the utilisation of underground water in power production and industry, and his analyses thus are not detailed and are expressed in accordance with technological conventions. However, as his Memoir includes the first map showing the distribution of various types of underground water in the Union of South Africa, it was thought advisable not to depart from his scheme of classification. It was though slightly modified to enable waters with total dissolved solids up to 200 mgm. per litre to be placed in group « E ».

V. NOTES ON THE WATER ANALYSES

The Tables on pages 228A to 228B contain twenty-five analyses representative of the various types of thermal water occurring in South Africa. Three are from deep, artesian boreholes, the remainder of springs. Several were made expressly for this paper, and no less than seventeen have not been previously published.* They have all been recalculated, when necessary, in terms of mgm. of ions per litre, and the millinormalities (N/1,000) have been worked out. This follows the recommendations of the International Society of Medical Hydrology (I.S.M.). One departure from the Society's recommendations has been the determination whenever possible of both CO_3'' and HCO_3'. A review of the literature showed that most chemists return the total carbonate as either CO_3'' or HCO_3'. Several of our thermal waters contain both (e.g. Badplaats, Chipise, and the alkaline Upper Karroo springs), and separate determination is essential in waters of fairly high pH. Figures for « total solids dried at 180°C. » are not given, for these are often unreliable unless the salts are all converted into sulphates.

The placd to be assigned to silica is problematical. Analysts return it variously as SiO_2, SiO_3'', H_2SiO_3'', $HSiO_3'$, etc.; but these are mere conventions and, as a rule, no definite reason can be given for the selection of one as opposed to another. The point is of great importance, for if the silica is present as SiO_3'' or $HSiO_3'$ it should be included amongst the anions.

From a study of recent chemical literature Roy[30] has concluded that at the concentrations characteristic of natural surface waters silica is probably present in the ionic condition. This is probably so at low concentrations only, of the order of 10 mgm. of silica per litre. Stumfer[34], for example, found that at concentrations of up to 5 mgm. per litre all the silica was molecularly dispersed, but that above this limiting figure there was an increasing proportion of « colloidal » silica. For thermal springs with rather high silica contents the dialysis experiments of Fresenius[12] suggest that the silica is probably present as di-silicic acid ($H_2Si_2O_5$), and that only a small fraction of up to 1 per cent. is colloidal. Unlike most chemists who have done research on the state of silica dissolved in water. Fresenius worked not on artificial solutions, but with actual spring waters having SiO_2 contents ranging from 33.2 to 180.2 mgm. per litre.

The present analyses show that the content of silica ,expressed as SiO_2, ranges from 11.0 to 86.0 mgm. per litre. In nearly all of the waters with a high pH value (i.e. the alkaline waters) the total millinormality of the anions exceeds that of the cations; silica can thus hardly be present as anions or the ionic balance would be completely disturbed. In the neutral or only slightly alkaline waters, on the other hand, there is generally a deficiency of anions and some of the silica may be present

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(*) Those credited to Messrs. W. Sunkel, P. Kok, C.F.J. van der Walt, C.J. Liebenberg and H.S. Schoeman were made in the laboratories of the Division of Chemical Services, Department of Agriculture. Those by Messrs. J. Gray, J. Muller and McLachlan and Lazar, consulting chemists, are included by kind permissino. The remainder were taken from Dr. Rindl's publications.[26]

in the form of ions. Unfortunately no acidic spring waters are know in South Africa. It is interesting to note that Dienert and Wandenbulcke[6], working with made-up solutions of sea water, found in neutral and alkaline solutions that the conversions of colloidal to non-colloidal silica is accelerated, but acids tend to prevent conversion. The conclusion reached was that in natural waters silica is almost exclusively non-colloidal, because the alkaline salts in the water would convert any colloidal silica to the non-colloidal form.

This line of argument is only valid if the cations are accurately determined. In many cases analysts arrive at their Na' figures by calculations based on the linking of positive and negative ions to form hypothetical salts. In several of the waters the Na' (and incidentally also the Mg'') was also determined gravimetrically by Mr. E. C. Haumann of the Division of Chemical Services, Pretoria. His results agreed remarkably closely with those previously obtained by calculation, as is shown by the following examples :—

Water	Na (calculated)	Na (gravimetrically)
Warmwaterberg	18.9 mgm. per litre	22.3 mgm. per litre
Zwartkops	141.8 » »	145.4 » »
Riffontein	234.7 » »	233.5 » »
T.G.I. (Trompsburg)	2,629 » »	2,665.3 » »

After a consideration of all these facts it was decided to express the total dissolved silica as SiO_2, the form in fact in which it was determined by the gravimetric method employed in the Government laboratories. To show the effect of regarding it as being present in the form of ions, millinormalities have been calculated as if it were present as SiO_3''.

The figures for dissolved hydrogen sulphide in the recent analyses (credited to W. Sunkel and P. Kok) are quite reliable, as special samples were taken in which it was immediately fixed as insoluble cadmium sulphide.

The chalybeatic waters deposited ferric hydrates in the bottles en route to Pretoria. Thus the pH values determined by analysts do not refer to the water as discharged by the springs. The iron and accompanying aluminium were returned as oxides by the analysts; in the Tables they are expressed as ions, but the millinormalities are enclosed in brackets to show that these items were not in solution when analysed. In the waters from Warmwaterberg, Fort Beaufort, Olifants valley, and Natal spa the total amount of iron and aluminium oxides did not exceed 4 mgm. per litre, and they were not returned separately. To arrive at an approximately figure for the millinormality it was assumed that half consisted of iron oxide.

Fluorine was determined in most of the waters. It ranges from mere traces up to 13.2 mgm. per litre. It is often present in fairly large amounts in borehole waters in South Africa[2], and its presence can no longer be assumed to indicate juvenile origin.

Of interest is the occurrence of 7.2 mgm. per litre of Sr'' in the water from a deep borehole near Trompsburg in the Orange Free State. It amounts, however, to only 0.09 per cent. of the total dissolved solids.

Saline residues obtained by evaporation were referred for spectrographic analysis to Dr. B. Wasserstein of the Geological Survey, whose report follows. The four waters from the Table Mountain series and the water from the spring at Fort Beaufort deposited iron and aluminium hydroxides before the samples reached the analyst. The deposits had to be removed from the bottles by acids and thus could not be included in the water taken to dryness. Such elements as Cu, Mn, Ba, and As would be precipitated together with the hydroxides, and for this reason the spectrographic data may be incomplete.

VI. SPECTROGRAPHIC DATA ON RESIDUES FROM THERMAL WATERS
BY B. WASSERSTEIN

The soluble solids obtained from evaporating the waters analysed by the Division of Chemical Services were subjected to qualitative spectrochemical analysis. The results are listed in the Table on page 212, and as a quantitative guide the intensity of each element-line has been visually estimated using an arbitrary scale from zero, to 10. This allows a comparison vertically, and to a lesser extent horizontally between the elements Pb, Mo, Li, Ba and possibly Mn, as their limits of detection are similar, viz. 0.002 per cent. or 0.003 per cent. of the solid. For Al the sensitivity is much less, perhaps not better than 0.3 per cent., whereas for Sr, it is much greater, probably 0.0001 per cent.—these two elements can certainly only be compared vertically in their respective columns.

In order to be able to judge the « order of magnitude » of the more important components, mixtures were synthesised with pure sodium chloride as a base—this was the commonest major component, by far, in the waters examined here. The mixtures made contained respectively 1 per cent.; 0.1 per cent; 0.01 per cent.; and 0.001 per cent. of the elements Pb, Mo, Li, Sr, and Ba. They were arced a number of times and the intensities noted for the different concentrations; these observations served as a basis for approximate quantitative estimation. It was found, for example, that an intensity of « 3 » corresponded to the presence of about 0.1 per cent. of Li, Mo, and Pb. For Ba, the same intensity denoted a slightly smaller amount but for Sr, a very much smaller amount, of the order 0.005 per cent.

Many more elements were sought than have been listed; some could not be found for reasons that cannot be elaborated here, and other omissions of results are due to possibilities of contamination; e.g. Zn was found in all the five waters from the Odendaalsrus area—no doubt derived from the galvanised lining of the boreholes. Very small amounts of Cr were found in most residues, but these results were discarded as experience has shown that chrome spatulas when used in scraping the residue from an evaporating dish, can cause contamination. In fairness to the chemists it must be stated that spectrochemical work was not envisaged. A few results on boron have been included, although at least some of this element was furnished by the glass containers.

The elements finally selected for examination are believed to show promise of special diagnostic significance. As a group their determination in waters is cumbersome by ordinary chemical methods and seldom undertaken—from this aspect alone information on such elements seemed desirable. The objective was not to seek out every trace element, but rather to concentrate on such as were capable of aiding geological interpretation—with this limited purpose in view the following method was found to be satistfactory.

The dried sample was placed in the cupped hollow of a copper anode of 5.8 mm. diameter; the cathode was similar but slightly pointed. A 240 v. D.C. source was used to run the arc for 20 seconds at 5 amps. A large Hilger Littrow-type spectrograph with a setting from 2800 A. to 5000 A. was available; for the few boron determinations a different setting and timing were used. As far as could be judged visually, reproducible results were obtained.

Although the investigation is exploratory rather than exhaustive, the results are encouraging, especially when one is mindful of the fact of the difficulty of matching composition of underground waters with their vagaries in geological formations. On the whole the spectrographic data underline similarities and differences that are apparent from the chemical analyses. The marked presence of Mn in the Table Mountain series waters, and the paucity of this element in the others, is not surprising in the light of our geological knowledge : small manganese ore bodies are known to occur in the Table Mountain series.

The identical results for the two Bushveld granite waters, although 50 miles (80 km.) apart, are what can be expected. These waters differ in respect of the selected elements from all the others with the possible exception of the Loubad one, which flows through rocks of the Waterberg system, i.e. through sediments formed partly of detritus from the Bushveld granites.

The five artesian waters from the Odendaalsrus area, viz. the boreholes V.K.1, V.K.3, J.R.1, K.1, and W.1 yield virtually identical results and this fact is in harmony with the geological data, for the waters traverse similar formations from a great depth. The boreholes are up to 10 miles apart and are a strong argument for the proposition that in respect of the selected elements, at least similar geological environment is reflected in similar composition of associated groundwaters.

The difference in composition between waters from a deep-seated and a shallow origin is shown by the results obtained for Fort Beaufort and Kat Valley; these springs are less than a half mile apart and issue from identical rocks of the Karroo system. The difference in Li-content, and to a lesser degree in Ba, is noteworthy. That between the shallow and deep-seated waters of Stinkfontein is not quite so marked.

Many more inferences can be drawn, but it is realised that more results are necessary before the influence of the host-rocks on the composition of thermal waters, even in a restricted sense, can be recognised. At this stage it seems reasonable

SPECTROGRAPHIC TABLE

LOCALITY	Solids mgm. per litre at 110° C	4,603⁑ Li	4,607 Sr	4,554 Ba	3,962 Al	4,031 Mn	3,903 Mo	4,058 Pb	2,498 B	Formation in which water occurs or struck.
Badplaats	335	4	8	tr.	2	1	2	2	n.d.	Archaean granite.
Tugela	1,518†	1	9	1	6	tr.	2	0	tr.	Archaean gneiss.
Riffontein	794	7	6	2	1	0	1	tr.	3	Rooiberg series.
Warmbaths	397	4	7	5	4	1	1	tr.	3	Bushvel granite.
Buffelshoek	459	4	7	5	5	1	1	tr.	5	Bushved granite.
Loubad	194†	5	7	6	1	tr.	tr.	1	5	Waterberg system.
Warmwaterberg	190	3	8	7	1	8	1	2	n.d.	Table Mountain
Olifants valley	156	2	7	5	3	9	tr.	tr.	n.d.	series.
Gamka valley	150	1	7	4	3	9	0	tr.	n.d.	» »
Zwartkops††	543	1	8	4	2	7	0	0	n.d.	Bokkeveld series (see p. 223)
J.R.1 — Jonkersrust††	3,724	4	9	6	4	1	0	tr.	n.d.	Ventersdorp lavas
V.K. 1 — Vermeulenskraal††	4.528	4	9	6	4	1	0	1	n.d.	(see p. 223)
V.K. 3 — Vermeulenskraal	4,544	4	9	6	5	2	0	1	n.d.	» »
K. 1 — Kaalpan††	3,800	3	9	5	5	1	0	tr.	n.d.	» »
W. 1 — Wolvepan††	3,996	4	9	7	5	1	0	tr.	5	» »
T.G. 1 — Trompsburg††	8,570	6	9	1	1	tr.	0	0	4	Pre-Karroo norite.
Fort Beaufort	528	6	7	1	2	0	tr.	0	n.d.	Beaufort series
Kat valley*	1,551	0	9	5	tr.	tr.	0	0	n.d.	(Lower stages)
Stinkfontein	754	7	6	tr.	0	1	0	1	4	» »
Stinkfontein**	859	5	8	tr.	1	tr.	1	1	1	» »
Kruidfontein††	995	7	8	3	2	tr.	2	0	3	» »
Kruidfontein**	1,397	5	10	1	6	1	tr.	0	tr.	» »

⁑ Wave-lenght. † Total sum of items determined. †† Borehole (thermal water). * Cold spring. ** Shallow borehole (cold water).

to assume that among the elements selected, some useful diagnostic ones have been included. An interesting by-product of this investigation has been the data—fragmentary though it be—on the elements Pb and Mo, whose geochemical cycles have received very little attention. The highest concentration of these elements (Badplaats) was of the order 0.005 per cent. of the solid, which is about 15 parts per 1,000 million in the water out of the Archaean granite for Pb and about 60 parts in the water at Tugela, out of the Archaean gneiss.

The highest concentrations of Li, Sr, and Ba are of the order of 0.1 per cent. in terms of solid, as far as could be determined by the technique employed; the amounts will be found to be not much less than 0.1 per cent. and certainly well below 1 per cent. when more accurate methods are employed—no greater claim to accuracy is made. The Riffontein « solid » may have a little more Li than 0.1 per cent. but the Trompsburg water will have the highest Li-content with approximately 9 parts per million. The highest Ba-content will probably be found in the Odendaalsrus waters, about half a part per million.

VII. SPRING GASES

In his well-known work « Ground Water, » published in 1937, C. F. Tolman states (p. 436) that spring gases consist chiefly of « carbon dioxide, hydrogen sulphide, and hydrocarbon gases. » Analyses of gas from South African thermal springs and thermal artesian boreholes do not support this contention. These gases fall naturally into two distinct classes.

One is characterised by predominance of nitrogen, and the gas may be regarded as air from which a varying proportion of oxygen has been abstracted and, as a rule, small amounts of carbon dioxide added. This carbon dioxide is probably either of atmospheric origin or due to the action of acids on calcareous rocks. The small amounts of methane present at Olifants valley and Badplaats is formed by the decay of organic matter in the spring pools. Springs emitting gas of this type occur in formations ranging from Archaean granite to the Devonian Table Mountain sandstone. Analyses, mostly new and previously unpublished, are given in the Table on page 214, and a few foreign ones have been added both for purposes of comparison and to show the widespread distribution.

Gases of the remaining class are marked by abundance of methane accompanied in the Aliwal North and Florisbad springs by hydrogen. Oxygen is either absent or present in very small amounts, but nitrogen commonly constitutes a large part of the gas. Without exception the springs rise in sedimentary rocks of the Karroo system (Upper Carboniferous to Triassic). Boreholes in this system occasionally strike gas that invariably proves to be composed dominantly of methane. Its origine is ascribed to distillation of coal of carbonaceous shales by the Jurassic dolerite intrusions. A similar origin may be postulated for the gases in the warm and tepid springs of the Karroo. Fairly thick layers of peat have accumulated around the Aliwal North and Florisbad springs, and it seems probable that there some of the gas has formed by decay of vegetable matter.

Although sulphuretted hydrogen taints the air around many South African thermal springs the amounts are too small to be determined by the ordinary methods of analysis.

Carbon dioxide is not an important constituent of any of the gases analysed. It does, however, constitute the bulk of the gas in small, cold springs issuing from fault zones in Dwyka tillite (Permo-Carboniferous) near Bizana. Farther north, at Bongwan, in the same Middle Cretaceous fault zone fairly large volumes of gas, consisting 99 per cent. of carbon dioxide, are exhaled. Gevers[13] concluded that the gas is not of volcanic origin but due to the action of sulphuric acid derived from oxidation on crystalline limestone.

VIII. RADIOACTIVITY

Rindl[26] quoted a few determinations of the radioactivity of thermal waters presumably measured by means of electroscopes :

Locality	Radioactivity in maché units (determined at source)	Geological formation
Zwartkops borehole	1.35	Cape system
Badplaats	6.5	Archaean granite
Machadodorp	12.3—12.8	Pretoria series
Malmesbury	73.3	Cape granite

NITROGEN-RICH GASES ASSOCIATED WITH SOUTH AFRICAN AND FOREIGN THERMAL WATERS.

Locality	Badplaats, Tvl.	Die Oog, Tvl.	Riffontein, Tvl.	Loubad, Tvl.	Warmwaterberg C.P.	Olifants Valley, C.P.	Zwartkops,CP. (borehole)	Buxton England,	Banff, Canada,	Leuker, Zwiterland.	Ranomafana, Madagascar.
Formation	Archaean granite.	Rooiberg felsite.	Rooiberg series.	Waterb. system.	TableMt. series.	TableMt. series.	Cape system.				
O_2	0.3	0.8	0.2	11	0.4	1.5	11.54	—	0.45	Nil	tr.
CO_2	0.2	—	0.7	2	2.7	3.5	3.43	0.88	1.11	0.9	0.55
H_2	nil	—	nil	—	nil	}3.9	*	—	—	—	—
CH_4	1.1	0.4	0.3	nil	nil		—	—	0.20	—	tr.
C_2H_6	nil	—	nil	—	nil	—	—	—	—	—	—
N_2+He+A, etc.	98.4	98.8	98.8	87	96.9	91.1	85.03	99.12	98.24	99.0	99.46
Totals	100.0	100.0	100.0	100	100.0	100.0	100.00	100.00	100.00	99.9	100.01
Date of analysis	1948	1939	1948	1941	1947	1947	1914	1882	c. 1918	c. 1943	c. 1926

* N_2 + He + Ne = 741 p.p. million.

COMBUSTIBLE GASES ASSOCIATED WITH SOUTH AFRICAN THERMAL WATERS.

Locality	V.K.I., O.F.S. (bore-hole).	Floris-bad, O.F.S.	Fort Beau-fort, C.P.	Cradock, C.P.	Tarka Bridge, C.P. (bore-hole).	Aliwal North, C.P.	Lake Mentz, C.P. (tepid).	Grasrand nr. Graaff-Reinet, C.P. (tepid)	Vreden-burg, nr. Marais Siding, C.P. (tepid)	Kruit-fontein nr. Eden-burg, O.F.S. (tepid)	Gruis-fontein nr Devon Tvl. (gas bo-rehole)
Formation	Ecca Series	Ecca Series	Beaufort Series	Beaufort Series	Beaufort Series	Beaufort Series	Dwyka Series	Beaufort Series	Beaufort Series	Beaufort Series	Ecca Series
O_2	0.4	—	—	1.1	0.1	3.2	tr.	0.3	1.2	—	—
CO	—	—	—	—		—	nil	nil	—	—	0.1
CO_2	nil	—	—	—	—	—	tr.	nil	nil	—	—
H_2	0.5	10.0	tr.	—	2.7	8.4	nil	nil	—	—	0.2
CH_4	80.8	71.5	100	83.4	94.0	55.2	39	98.5	33.3	79	81.2
C_2H_6	—	—	—	—	—	—	nil	nil	—	tr.	2.1
N_2 + He + A, etc.	18.3	18.5	tr.	15.4	3.2	33.2	60	1.2	65.5	20	14.7
Totals	100.0	100.0	100	99.9	100.0	100.0	99	100.0	100.0	99	99.2
Date of analysis	1947	1916	1948	1946	1912	1927	1945	1945	1942	1944	1935

Analysts : Zwartkops — J. Dewar; Die Oog and Loubad — C.F.J. van der Walt; Florisbad — M.M. Rindl; Aliwal North — Messrs. Littlejohn and Whitby; Tarka Bridge — J. Moir: Remainder — Fuel Research Institute of South Africa, Pretoria.

The activity is due to the presence of radon, and in the case of the Malmesbury spring water it fell to between 3.5 and 3.6 maché units after only three hours.

Brandvlei scalding spring was investigated by Mr. J. Muller, Consulting Chemist, during 1946. The water showed 50 to 53 maché units, and the spring gas 450 maché units per litre. This spring rises in the Cape system, and the radom content is rather high for a water of this type.

A large sample of water from the Badplaats scalding spring was taken to dryness and the residue submitted to Dr. D. J. Simpson of the Geological Survey. Examined with beta-ray Geiger-Muller counters it showed only traces of radioactivity probably due to the 1.2 per cent. potassium present. No increase in activity was noted after the sample had been stored for two months.

IX. SPRING DEPOSITS AND AGE OF SPRINGS

Very few South African thermal springs have given rise to deposits. The only ones known of any importance are ochres and ferruginous and manganiferous sinters deposited by springs issuing from the Table Mountain sandstones in the belt of fold mountains along the south coast of the Cape Province. Fairly extensive deposits occur around Caledon, Warmwaterberg, Toverwater, and Amanzi[12] springs. The Olifants valley and Gamka valley springs are also markedly chalybeatic, but as they issue on the banks of rivers and are thus often inundated, there has been no opportunity for sinters to accumulate.

In the deep Zwartkops borehole, near Port Elizabeth, scalding hot water was struck at depths of between 3,400 and 3,560 feet (1,036 m.—1,085 m.) in presumably Bokkeveld sediments;[28] the water, however, is believed to be derived mainly from the underlying Table Mountain sandstones. Ochre is deposited in the overflow channels.

Analyses show that the manganese content of these ochres and sinters increses westwards from Zwartkops to Caledon, and it is of interest to note that still nearer to Cape Town small deposits of psilomelane occur in the Table Mountain sandstone. It is possible that some of them mark the sites of extinct springs.

Ochre is also deposited by numerous cold springs rising from Table Mountain sandstone. An analysis of such an ochre from a spring in the Gamka valley has been added for comparison. The manganese content is much lower than in thermal water deposits.

The phosphate content of all is fairly high, and an appreciable amount of barium was found in the Warmwaterberg sample.

Spring	Caledon	Warm-waterberg	Zwartkops	Gamka Valley (cold spring)
Material analysed	Average sample from mound	Recently deposited ochre	Recently deposited ochre	Recently deposited ochre
SiO_2	5.91	7.8	6.74	2.8
Al_2O_3	2.21	n.d.	n.d.	n.d.
Fe_2O_3	50.72	57.5	70.9	67.1
MnO	—	0.55	0.50	0.16
MnO_2	27.20	8.55	1.36	nil
MgO	n.d.	4.8	0.3	tr.
CaO	n.d.	0.8	0.7	0.2
BaO	n.d.	1.45	n.d.	n.d.
P_2O_5	2.31	1.18	4.20	2.67
H_2O+	9.47	9.59	10.0	13.2
H_2O-	1.38	7.64	6.0	13.9
	99.20	99.86	100.60	100.03

Analysts : Caledon and Zwartkops — C. F. J. van der Walt; Warmwaterberg and Gamka River — E. C. Haumann. Division of Chemical Services, Pretoria.

According to Dr. L. G. Boardman of the Geological Survey the Caledon sinter mound would weigh about 1,000,000 tons (907,000 metric tons). The one at Warmwaterberg is estimated to contain about 600,000 tons (544,000 metric tons).

Assuming that the deposition of iron and manganese is complete, and that the composition and flow of the waters have remained constant, and that there has been no erosion or solution of the deposits, the Caledon and Warmwaterberg sinters have respectively taken approximately 300,000* and 850,000 years to accumulate. It is of course, most unlikely that conditions remained uniform, but these figures provide a rough indication of the order of time involved. On geological grounds it is probable that hot water has issued from these springs ever since the present topography was developed during hte Late Tertiary.

These age estimates may be compared with Meinzer's statement[36] that most of the thermal springs in Nevada, Utah and southern Idaho came into existence before the close of the Tertiary. The Yellowstone Park springs, on the other hand are probably only some 20,000 years old, and those of Iceland between 10,000 and 50,000 years[36].

X. THERMAL ARTESIAN BOREHOLES

Several deep boreholes have struck thermal artesian water. In one, near Port Elizabeth, the water is scalding; in another, near Trompsburg, it is hot; in three near Odendaalsrus it is warm.

Borehole	Latitude °S.	Longitude °E.	Depth at which thermal water struck Ft.	M.	Formation in which thermal water struck	Temp. °C at surface	Flow Imp. galls / 24 hrs.	Cu. metres / 24 hrs.
Zwartkops[28],[29]	33°53′	27°37′	3,400	1,036	Bokkeveld series(?)	54.5	250,000	1,134
T.G.I.* ...	30° 3′	25°44′	4,700	1,433	Pre-Karroo norite	37.2	24,000	109
J.R.I.* ...	28° 6′	26°44′	3,560	1,085	Ventersdorp series (pre-Cambrian)lavas	34.0	48,000	218
V.K.1* ...	28° 6′	26°49′	2,560	780	Ventersdorp series lavas	32.7	48,000	218
W.1*	28° 2′	26°41′	—	—	» » »	28.8	60,000	272

* This information was kindly supplied by the mining companies concerned. The temperature of the artesian waters in the Odendaalsrus area were taken by Dr. L. T. Nel of the Geological Survey.

In the Odendaalsrus area three other recently drilled holes have struck tepid water at comparable depths. Gas consisting dominantly of methame is discharged from them all; it enters the boreholes from the Permian Ecca series, in J.R. 1, for example, at a borehole depth of 980 feet (299 m.).

The borehole at Zwartkops has been developed into one of the leading spas of the country, and the J.R. 1. and V.K. 1. waters are used locally for medicinal purposes.

Several boreholes have been drilled near thermal springs with the object of obtaining more water. On the farm Tugela, on the Limpopo river in the northern Transvaal, a borehole 65 feet (20 m.) deep, drilled in 1947, gives a flowing supply of 15,600 gallons (71cbm.) of hot water per day. The natural spring 100 feet away has a discharge of only 2,650 gallons (12 cbm.) a day, and according to the owner its flow has not as yet been appreciably affected by the artificial release of pressure. The borehole water has a temperature of 48.9°C. whereas the maximum temperature measured in the spring was only 42.8°C. A borehole about 100 feet (30 m.) deep at Warmwaterberg flows at the rate of 31,000 gallons (141 cbm.) a day, but the water although warm is 17.7°C. cooler than that of the nearby hot spring.

Former warm springs at Kruidfontein and Amanzi have dried up as a result of heavy pumping of boreholes in the immediate vicinity.

Boreholes only 65 to 225 feet (20-69 m.) deep at Tarka Bridge[35] yield warm (about 27°C.) artesian, sulphuretted water accompanied by inflammable gas, mainly

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(*) A sample of water from the Caledon spring kindly collected by the proprietors in May, 1948, was partially analysed by Mr. E.C. HAUMANN of the Division of Chemical Services, Pretoria. The results were :

SiO_2	53·3	mgm. per litre	
Fe_2O_3	4·1	» »	(as $Fe^{\cdot\cdot}$ 2·9)
Mn_3O_4	4·6	» »	(as $Mn^{\cdot\cdot}$ 3·3)

On the basis of the respective manganese oxide contents, the mound is 180.000 years old; whereas when the values for ferric oxide are used the age works out at 370,000 years.

methane. The holes are of particular interest as they are tidal, although situated some 160 miles (255 km.) from the sea. Here too, weak thermal springs have been dried up.

In the Karroo many other shallow boreholes have struck sulphuretted water, which is generally tepid indicating a fairly deep-seated origin, probably from blind fissures.

XI. THERMAL MINE WATERS

There are a few published records of thermal water having been struck in the deep gold mines of the Witwatersrand.

A narrow, water-bearing fissure in a dyke was intersected at a depth of 2,680 feet (817 m.) in a shaft on the Randfontein Estates mine. The water had a temperature of 25°C., which was the same as that in the main workings at this depth. On analysis the dissolved solids amounted to 134 mgm. per litre, the water being characterised by predominance of SO_4'' and virtual absence of Cl'.[27] The supply of water was weak, but on the East Rand Proprietary Mines a supply of about 1,800,000 gallons (8,166 cbm.) of warm water per day was struck in a fault zone in Witwaterstand quartzites. Temperature readings at various points in the fissure system at vertical depths of between 5,261 and 6,056 feet (1,603-1,846 m.) ranged between 33.3° and 35°C., the rock temperatures being about 1°C. lower. The water was sulphuretted, containing 10 mgm. per litre of dissolved hydrogen sulphide, and at first was very soft dissolved solids amounting to only 160 mgm. per litre. As dewatering proceeded the concentration of salts in the water increased to approximately 1 gm. per litre, but the dominant constituents remained the same, viz. $NaHCO_3$ and Na_2CO_3.[1]

Thermal water has also been struck in the copper mines at Messina, but no details are available.

XII. THE RELATIONSHIP BETWEEN GEOLOGICAL FORMATION AND CHEMICAL COMPOSITION OF SOUTH AFRICAN THERMAL WATERS.*

(*a*) Archaean Granites, Gneisses, Amphibolites, etc.

In these most ancient rocks the thermal springs are found along the Orange river in the Cape Province, in Natal, and in the eastern and northern Transvaal. Most of them are hot, and two—Badplaats and Tugela valley—are scalding. The flows range up to about 180,000 gallons (817 cbm.) a day, and as usual the strongest springs are hottest.

The three springs in the Orange river valley have not as yet been analysed. The remainder are of three types characterised by dominance of chlorides, sulphates, and carbonates, respectively. Those having NaCl as the main constituent occur in the northern Transvaal and contain on the average 1 gm. of dissolved solids per litre. Those with a high SO_4'' content are Tugela in the extreme north of the Transvaal and Lilani and Tugela valley in Natal. Soda-rich alkaline waters occur at Natal Spa, Badplaats, and Paddysland. In all save Paddysland $Ca^{\cdot\cdot}$ considerably exceeds $Mg^{\cdot\cdot}$. Silica is a prominent constituent especially in the soda-rich alkaline waters.

It is interesting to note that Bond's[2] work shows that the water from shallow boreholes in the Archaean granites ,etc. may be divided into the same three types, or, in other words, the thermal springs discharge water essentially similar in composition to the normal underground supplies in the vicinity. The only important point of difference is the ratio of $Ca^{\cdot\cdot}$ to $Mg^{\cdot\cdot}$. In the non-thermal waters, save in Natal, $Ca^{\cdot\cdot}$ is subordinate to $Mg^{\cdot\cdot}$, whereas in nearly all the thermal waters the reverse is the case. The exception, the spring at Paddysland in the northern Transvaal, is barely thermal, having a temperature of only 26°C. It is possible that at depth the thermal waters extract the $Ca^{\cdot\cdot}$ from calcium-magnesium silicates, whereas the Mg is retained as relatively insoluble hydrated silicates.

(*b*) Pongola System

At Sulphur Springs near the Swaziland border the warm water apparently issues from rocks of this Proterozoic system. The analysis shows very weak mineralisation, with sodium carbonate as the main constituent. The water is quite strongly alkaline and contains a high proportion of silica.

(*) For representative analysis, refer to Tables.

There do not appear to be any published analyses of borehole waters from the Pongola system. A cold, sulphuretted spring from Pongola quartzites was found to contain 253 mgm. per litre of dissolved solids, the predominant constituents being $NaHCO_3$ and $NaCl$. Like the thermal spring its pH value is high, being 9.55.[19]

(*c*) Pretoria Series

The warm spring at Machadodorp and those at Badfontein and De Bad farther to the north emerge from rocks of the Pretoria series of the pre-Cambrian Transvaal system. Only the Machadodorp water has been analysed. Although classed by Rindl[26] amongst the sulphur springs the analysis shows less than 1 mgm. per litre of dissolved hydrogen sulphide. The water is not strongly mineralised, and the characteristic constituents Cl', HCO_3', and SiO_2 are present in almost equal quantities. The SO_4'' content is low, and $Ca^{\cdot\cdot}$ exceeds $Mg^{\cdot\cdot}$ four times. Apart from the $Ca^{\cdot\cdot} : Mg^{\cdot\cdot}$ ratio the water is quite typical of that from shallow boreholes in the shale zones of the Pretoria series; in these the $Ca^{\cdot\cdot} : Mg^{\cdot\cdot}$ ratio is almost exactly 1 : 1.[2]

(*d*) Rooiberg Series

This pre-Cambrian series embraces both felsitic lavas and sediments. The hot waters at Die Oog and Welgevonden, which rise from a fault in the felsites, are of low concentration, the characteristic constituents being $NaHCO_3$ and $Ca(HCO_3)_2$, with subordinate $NaCl$. The warm water issuing from a fault in Rooiberg series quartzites at Riffontein, on the other hand, is more than three times as concentrated, and $NaCl$ is the major constituent followed by $NaHCO_3$. The Rooiberg series thermal waters are marked by great excess of $Ca^{\cdot\cdot}$ over $Mg^{\cdot\cdot}$.

(*e*) Bushveld Granites

The thermal sgrings at Warmbaths, Buffelhoek, Grovesbad, Goederede, and Vischgat, all in the central Transvaal, rise in Bushveld granite. In temperature they range from warm to scalding and in flow from weak to about 160,000 gallons (726 cbm.) a day.

Only the Warmbaths and Buffelhoek waters have been analysed. Both chemically and spectrographically they are remarkably alike. Chemically they are quite typical of Bushveld granite borehole waters but are somewhat more concentrated, more siliceous, and have rather higher sodium chloride and sodium carbonate contents. The main point of difference is the $Ca^{\cdot\cdot} : Mg^{\cdot\cdot}$ ratio, which is about 6 ½ : 1 for the thermal waters and 2 : 1 for the cold borehole waters.

(*f*) Cape Granite

The warm spring at Malmesbury rises from a fault zone in Cape granite. The water is saline, having $NaCl$ as chief constituent. On the contrary, borehole waters are of low concentration but also marked by predominance of $NaCl$.

(*g*) Waterberg System

This system which in the central Transvaal is made up almost entirely of a thick succession of sandstones may be of Late pre-Cambrian age. The warm springs at Loubad in the Central Transvaal represent deep artesian water from the basal layers of the system. The water is of low concentration, the main constituent being $Ca(HCO_3)_2$. Cl' is low and SO_4'' virtually absent. The non-thermal waters from cold springs and boreholes in this system have essentially the same composition but are of still lower concentration and contain proportionally more $Mg^{\cdot\cdot}$.[18]

(*h*) Cape System

In the western Cape Province the thermal springs all rise on the flanks of mountains formed by Mesozoic foldings. The aquifer is the Devonian Table Mountain sandstone, and most of the springs issue from fault planes. The flows are fairly strong, and with the exception of Gamka valley the water is hot or scalding. Analyses, however, show that save for a much lower content of SiO_2 this Gamka valley warm spring and the nearby Olifants valley scalding spring discharge waters of almost identical composition.

Chemically the waters are all of low concentration, the total dissolved solids ranging from 77 mgm. per litre at Goudini to 190 mgm. per litre at Warmwaterberg. In the west, at Goudini and Brandvlei, SiO_2 is the most abundant constituent followed as regards the anions by HCO_3', Cl', and SO_4''. At Caledon, which is much nearer the sea, Cl' exceeds HCO_3'. At Warmwaterberg and Olifants valley HCO_3' and not SiO_2 is the major constituent, whereas at Toverwater Cl' exceeds both. According to Bond[2] this latter would be typical of the non-thermal waters of the Table Mountain series, in which the order of abundance is Cl', HCO_3', SiO_2 and SO_4''. The general predominance of SiO_2 in the thermal waters is doubtless due to its greater solubility in hot water.

Several of the springs are chalybeatic and deposit hydrated oxides of iron and manganese. The variations in iron and manganese content were discussed in the section dealing with the spring deposits.

The hot water struck between depths of 3,400 and 3,560 feet (1,036-1,085 m.) in the borehole at Zwartkops is more concentrated than that from thermal springs in the Cape system, and contains proportionally less HCO_3'. The warm water from the Amanzi boreholes, also near Port Elisabeth, is even more concentrated and characterised by considerable excess of SO_4'' over Cl', fairly high $Fe^{\cdot\cdot}$ and $Al^{\cdot\cdot\cdot}$ content, and absence of carbonates. The Zwartkops waters is mixed of origin; probably Table mountain series, Bokkeveld series, and the base of the Cretaceous. The Amanzi water is also a mixed Table Mountain series and Cretaceous water.

(*i*) Karroo System

Sediments of this system, which ranges from Upper Carboniferous to Rhaetic in age, cover the central part of the Cape Province, nearly the whole Orange Free State, and large areas in Natal and the Transvaal. It is divided into the following series :—

Stormberg series
Beaufort series
Ecca series
Dwyka series

All the thermal springs in the Karroo system are only warm and vary in strength from flows of only a few thousand gallons up to 840, 600 gallons (3,813 cbm.) a day at Aliwal North. Most of them rise alongside dolerite dykes of early Jurassic age.

Only the Black Umfolosi spring in Natal actually issues, in this case tillite at the surface from rocks of the Dwyka series.

The springs in the eastern Orange Free State emerge from rocks of the Ecca series. The waters are rich in $NaCl$, and the total dissolved solids exceed 2 gm. per litre. Such a high salt content is not characteristic of the underground water in the Ecca series but rather of the underlying Dwyka series, through rocks of which it seems probable the thermal waters have percolated. Deep boreholes to the south of Odendaalsrus and near Trompsburg, both in the Orange Free State, have also struck thermal waters characterised by predominance of $NaCl$. As already mentioned the waters were struck in lavas of the pre-Cambrian Ventersdorp system and in norite of presumably Bushveld Complex age, but it is probable that much of the $NaCl$ was derived as a result of downward percolation through the Dwyka series- which overlies these ancient rocks. This seems a reasonable supposition, for Bond's[2] analyses show that the waters in Ventersdorp lavas and Bushveld norite are charac, terised by a rather low Cl' content and are temporary hard carbonate waters.

In the central and eastern parts of the Cape Province the thermal waters in the Beaufort series sediments nearly all have $NaCl$ as the predominant constituent, with total dissolved solids ranging from 496 mgm. per litre at Badsfontein to 754 mgm. per litre at Stinkfontein. From the analyses given by Bond[2] it is evident that the shallow underground waters from the Lower beds of the Beaufort series are saline, whereas a high sodium bicarbonate content characterises the waters from the Middle and Upper beds. The difference is believed by Bond to mark a transition from an estuarine to a fresh-water origin. The $NaHCO_3$ in the upper beds is considered to be due to a higher proportion of unkaolinised feldspar[2].

The thermal springs at Rooiwal and Cradock, however, have CO'' and HCO_3' as the chief anions, accompanied by a fair amount of SiO_2. Ca'' is very low but greatly exceeds $Mg^{\cdot\cdot}$. Near Rooiwal another thermal spring, Badsfontein, issuing from about the same horizon in the Middle Beaufort stage and having almost exactly the same temperature and flow, is saline. Thus it is probable that Badsfontein is a deeper-seated spring but that much of the excess heat has been lost before the water emerges at the surface. The difference in composition between the Cradock and not very far off Tarka Bridge thermal waters can be similarly explained.

The warm water at Stinkfontein, although issuing from Lower Beaufort strata, is characterised by a high $NaHCO_3$ content and the presence of much SO_4''

s well as Cl'. $\mathrm{Mg}^{\cdot\cdot}$, present to the extent of 15 mgm. per litre, greatly exceeds $\mathrm{Ca}^{\cdot\cdot}$; judging by the analyses in Bond's memoir such preponderance of $\mathrm{Mg}^{\cdot\cdot}$ is most unusual in Karroo waters.

In the extreme north of the Transvaal the scalding Chipise spring issues from Upper Karroo beds overlying Archaean gneiss. According to Söhnge[31] the point of emergence is determined by a dolorite dyke approaching a fault and the contact of Cave sandstone with a dolerite sill. Mr. H. N. Visser who has recently resurveyed the area is, however, of opinion that the water issues at the intersection of two faults. Movements have taken place along faults at intervals up to post-Permian time. The course of the water has not been determined, but it is probably of mixed Archaean granite, Waterberg system (where the intake area may lie),and Upper Karroo origin It is moderately mineralised, the main constituents being NaCl and $\mathrm{NaHCO_3}$. The $\mathrm{SiO_2}$ content is fairly high, but $\mathrm{Ca}^{\cdot\cdot}$ and $\mathrm{Mg}^{\cdot\cdot}$ are present in very small amounts, the former greatly exceeding the latter.

XIII. SOURCE OF THE WATERS AND THEIR HEAT

The youngest igneous rocks known in the Union of South Africa, the intrusions of kimberlite and melilite basalt, date back to the last stages of the Cretaceous. Admittedly the crater of the Pretoria salt-pan has generally been considered to be an extinct volcano of camparatively recent age, but the latest investigations have suggested that it is probably a meteor crater. This complete absence of Recent or Late Tertiary igneous activity seems to rule out eruptive rocks as a source of the heat of the South African thermal springs, or of their waters or gases. This is assuming that there are no relatively young abyssal intrusions.

It has been found possible to explain the origin of all the thermal springs that have been investigated in detail by structures permitting water of meteoric origin to descend to depth, take up earth heat, and return to the surface at such a rate that much of this heat is retained. Some geothermal surveys have been made in South Africa, and when the temperatures at the depths to which the water is assumed to descend in such structures are calculated the results agree closely with the observed spring temperatures. Thus it is not necessary to postulate the presence of juvenile water, but the possibility cannot be ruled out that such water may contribute to the discharge of some of the springs. Connate water, on the other hand, almost certainly forms part of the thermal water flowing from the deep boreholes recently drilled in the Odendaalsrus and Trompsburg areas of the Orange Free State and may be present in some of the springs as well.

Geological sections illustrating the structures determining a few representative thermal springs form figure 2. One of the simplest is the artesian basin around Loubad in the central Transvaal[18]. There rain falling on moderately high ground is believed to percolate through sandstones of the Waterberg system (late pre-Cambrian ?) until the basal shales or the impervious felsites of the Rooiberg series are reached. The water emerges from fractures next to a sharp fold that determines the south-west end of the basin. The maximum depth in the basin is about 3,600 feet (1,097 m.), and as the geothermal step is about 220 feet per 1°C. the water would be heated to a temperature of about 36°C. This agrees very well with that measured in the main orifices of the springs, viz. 31.7° to 34.1°C.

The small warm spring at Buffelhoek, also in the cental Transvaal, rises from Waterberg sandstone next to a diabase dyke. The water, however, is chemically akin to that in granites, and not to that in sandstones. Geological mapping of the vicinity showed that the water in all probability percolates in Bushveld (Red) granite forming an artesian slope beneath the Waterberg sandstones until the barrier fromed by the dyke is reached.[20] From the geological section it appears that the water would descend to a maximum depth of about 3,000 feet (915 m.), where the rock temperature would be about 31°C. In the eye of the spring the temperatures is 30.5°C. It is of interest to note that chemical and spectrographic analyses showed that the water is almost identical in composition with that issuing from a fault in Bushveld granite at Warmbaths fifty miles (80 km.) to the east-south-east.

The spring at De Bad, in the western Cape Province, is situated in a deep valley. As will be seen from the section its origin is relatively simple. Rain falling on a mountain range formed by sandstones of the Table Mountain series (Devonian) descends in a deep, fairly sharp syncline and emerges behind a barrier formed by a prominent shale band.

The hot spring at Goudini, also in the western Cape Province, rises from a mid-

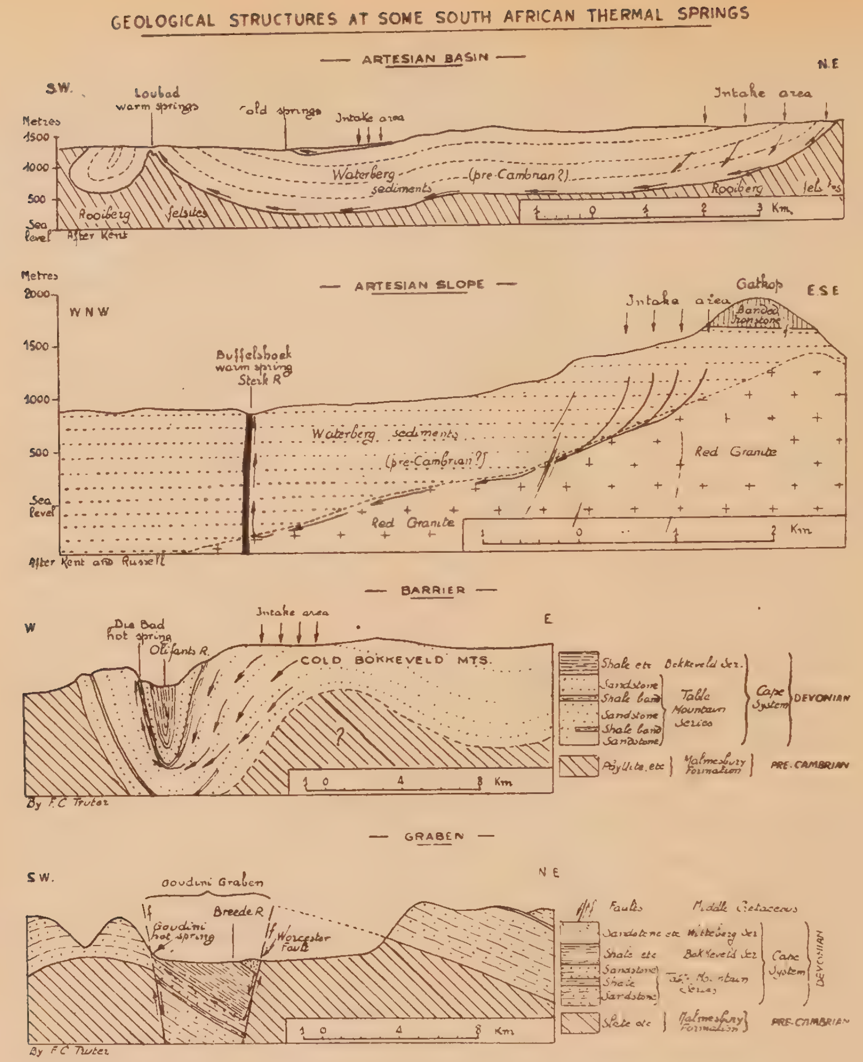

FIG. 2.

Cretaceous fault in the Table Mountain series. According to Dr. F. C. Truter of the Geological Survey, who investigated the area in 1938, the water probably percolates along and down the plane of the Worcester fault, which is one of the major faults of South Africa. This fault traverses mountainous country, and pressure must thus be built up in the water descending in the fracture zone. This water probably ascends in the sandstone confined between a shale band and the overlying shales of the Bokkeveld series until the plane of the Goudini fauls is encountered, where the pressure is relieved. Some of the hot water may originate from rain percolating in the Goudini fault zone.

Recent detailed mapping by Mr. D. P. van Rooyen of the Geological Survey showed that both faulting and an intrusion of diabase are responsible for the emergence of hot water at Welgevonden in the north-central Transvaal. Rain water is assumed to percolate through sandstones of the Waterberg system until a diabase dyke is reached. This dyke eventually turns and follows an almost vertical fault cutting through felsites of the Rooiberg series (pre-Cambrian). The hot water rises in the fault zone alongside the dyke.

The shallower thermal artesian boreholes do not call for any special comments. The deep Zwartkops borehole was drilled in an artesian basin formed of Lower Cretaceous rocks lying unconformably on what was taken to be the Bokkeveld series.[28] It is less easy to account for the flowing supplies struck at great depths in the Orange Free State. Compaction of the aquifer, lavas of the Ventersdorp system around Odendaalsrus and pre-Karroo norite at Trompsburg, seems the most likely explanation.

Together the seven South African scalding springs bring to the surface approximately 150,000,000 kg. cal. of heat per 24 hours. The information regarding the temperatures and flows of the remaining thermal springs is not complete, and reliable calculations of the discharge of heat cannot be made. It is estimated that the total for all the thermal springs in South Africa is about 600,000,000 kg. cal. per 24 hours or 5.8×10^{-10} cal. cm^2. sec. averaged over the whole surface of the country. This may be compared with Bullard's[4] figure of 1.16×10^{-6} cal. cm.^{2} sec. for the mean heat flow in South Africa. Thus the heat brought to the surface by thermal water is only one two-thousandth part of that normally dissipated by conduction through the crust.

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PART II. THE THERMAL WATERS OF SOUTH WEST AFRICA

I. Introduction

With an area of 82.292,180 hectares (317,725 sq. miles) the Territory of South West Africa is about three-quarters of the size of the Union of South Africa. In the Territory twenty-four springs are known to be thermal, and on the basis of temperature they may be classified as follows :—

Warm	8
Hpt	4
Scalding	9
Temperature unknown	3

As the mean annual temperature in the mountainous tract around the capital, Windhoek, is 19°C. (66.2°F.), the same figure has been used as in the Union to divide thermal from non-thermal springs, viz. 25°C.

The positions of the various springs are shown in the map, fig. 3, which is based on all available published information supplemented by a personal communication from Dr. H. Martin, geologist attached to the Administration of the Territory.

Relatively little is known about these thermal springs. Many of them occur in valleys in the central highlands between Rehoboth and Omaruru. On this mountainous tract, which stands some 1,500 to 2,000 metres (4,921 to 6,562 ft.) above sea-level, the rainfall is fairly high, 356 to 406 mm. (14 to 16 ins.) a year. Westwards it falls rapidly to less than 25 mm. (1 in.) in the coastal Namib desert. In the far north the coastal belt passes over from a sand desert into a fairly mountainous region known as the Kaokoveld. Here the annual rainfall is about 76 to 152 mm. (3 to 6 ins.) a year, and there are at least four springs reputed to be thermal.[6] West of the Kaokobeld, in the Otavi Mountain Land, where the rainfall is about 560 mm. (22 ins.) a year, there are several fairly strong springs, only two of which, those at Rietfontein and Naumotoni, are even lukewarm. In the extreme south of the Territory, in the basins of the Orange and Fish rivers, the rainfall is very low, about the same as in the Kaokovled; but from this arid region no less than six thermal springs are known.

II. Temperatures of the Springs

Compared with the Union a much higher percentage of the springs are scalding. Doubtless this is due, in part, to the territory being less well-known, with the result that many lukewarm springs have not attracted any attention. Furthermore, a deep-seated spring in arid regions has a better chance of survival than a shallower one. Finally, the central highlands of South West Africa are highly faulted, and minor earthquakes show that movements still take place along several of the fault planes. By thus keeping fissures open deeper penetration of water is made possible.

The following temperature observations have been made :—

	°C	Observer
Gross Windhoek (Pahl spring)	79.8	Gevers, 1931.
Omburu	76.5	Rehbock, 1896-97.
Gross Barmen	65	» »
Klein Barmen	61	» »
Omapyu	61	Schinz, 1885.
Aiais	55	Range, 1906-08.
Klein Windhoek	54.2	Gevers, 1931.
Rehoboth	52	Rehbock, 1896-97.
Near Outjo (position unknown)	46	Cock, 1929.
Ganikobis	40	Range, 1906-08.
Warmbad	37.5	Range, 1906-08.
Sneyrivier	28	H. Martin, 1948.
Rietfontein	27.8	Owners, 1948.

III. Discharges of the Springs

At Gross Windhoek scalding hot water formerly emerge at six main springs and a number of minor ones, all strung out over a lenght of 1 ½ km. along a volca-

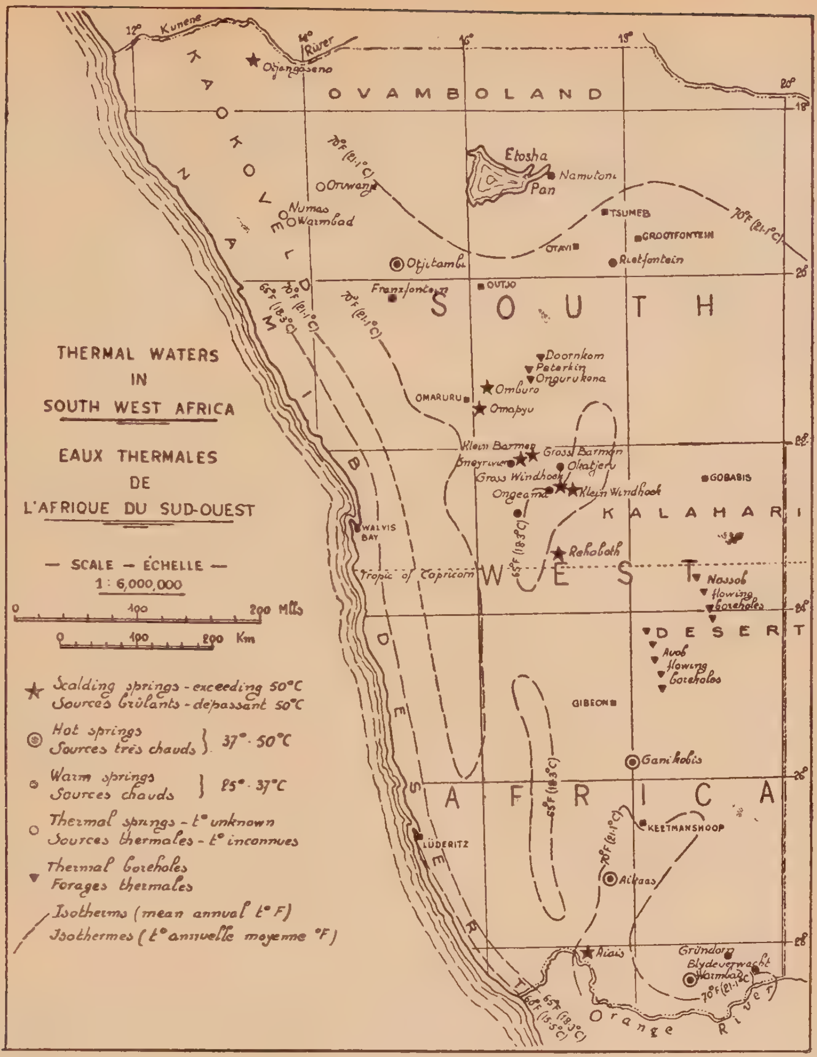

Fig. 3.

nic fissure. The hottest, the Pahl spring, had a discharge of about 400 cbm. (88,200 galls.)[3] in 1923. As the supply was insufficient for the growing needs of the town, which depended solely on the thermal springs, several boreholes were sunk, and as a result of the increased withdrawal of water all the former natural springs have dried up. In 1931, according to Gevers[3], the combined yield of all these boreholes was 2,499-2,817 cbm. (550,900-621,000 galls.). At Klein Windhoek, 3 km. to the east the natural springs have also become extinct as a result of the tapping of the supply by numerous boreholes.

According to Dr. H. Martin the Gross Barmen spring yields about 720 cbm. (159,000 galls.) of water a day. At the other end of the Territory the warm spring at Blydeverwacht was estimated by Haughton and Frommurze[4] to have a discharge of about 180 cbm. (40,000 galls.) a day. Apparently the only other that has been gauged is the lukewarm one at Rietfontein, which, according to the owners, the Rietfontein Co-operative Creamery, Ltf., yields about 2,140 cbm. (472,000 galls.) per day.

IV. Geology of the Springs

As far as is known most of the thermal springs, and especially the hotter ones, rise in fault or fracture zones. The Windhoek springs, for example, rise in almost vertical fissures associated with volcanicity possibly of Cretaceous age. The lukewarm ones at Okatjeru and Ongeama (Okanjama) also lie on associated breccia-filled fissures.[3] Gross Barmen emerges at the north end of a fault that passes southwards into the Khomas Highlands.

The lukewarm Rietfontein spring, on the other hand, issues at the contact of the Otavi limestones (pre-Cambrian) and Archaean schists.

According to Dr. Martin the Sneyrivier and Ganikobis springs rise alongside dykes. At Sneyrivier the country rock is Archaean granite, whereas at Ganikobis it is the basal sediments of the Dwyka series (Carboniferous-Permian).

Blydeverwacht is a barrier spring, the barrier being a reef of amphibolite cutting across sheared gneisses[4].

V. Thermal Boreholes

Warm water has been struck in many of the boreholes in the artesian areas to the north-east of Gibeon. The main aquifer, a fairly thick sandstone in the Permian Ecca series, is struck at depths of between 43 and 166 m. (140 to 545 ft.). The strongest hole yielded 4,083 cbm. a day (900,140 galls.). The temperatures of the water taken at two boreholes in the Auob area were 32°C. and 33.7°C.[2]

Boreholes recently drilled in north-south trending fault zones north-east of Omaruru have also struck warm water. The following particulars were obtained through Dr. Martin :—

Borehole	Depth		Water-level from surface		Yield–24 hrs		Temp. °C.
	m.	ft.	m.	ft.	cbm.	galls	
Doornkom	81	265	33	108	>327	>72,000	35—40
Peterkin	83	274	32	106	>390	>86,000	35

It should be noted that in South West Africa water in shallow boreholes generally has rather a high temperature, about 23° to 25°C., for during the day the barren rocks and soil acquire a far higher temperature than the air.

VI. Composition of the Waters

Only the thermal waters at Windhoek[3], Gross Barmen,[5] Warmbad,[5] and a spring near Outjo[1] have been analysed. The first two rise from Archaean schists, Warmbad from granite-invaded gneisses, while the exact locality of the Outjo spring is not known. According to Cock[1] it also rises in granite.

In his paper on the Windhoek springs Gevers[3] included several analyes. The most complete of these are included in the following Table. They are of water from the Pahl spring borehole and from the Jan Jonker boreholes, 2 km. south of the Pahl spring, but on the same fissure.

The analyses have all been recalculated on an ionic basis.

The Windhoek and Gross Barmen springs are only weakly mineralised, whereas Outjo and Warmbad are saline. Compared with the thermal waters in the Union the most striking feature is the high potassium content, particularly at Gross Barmen and the Jan Jonker boreholes, where potassium exceeds sodium. This is doubtless due to the higher proportion of potassium in the Old granites and schists of South West Africa.[5]

Sulphate is also high in the analysed waters and at Windhoek is far more prominent than chloride. The pH values were not determined, but for the sake of uniformity all the carbonate has been recalculated as bicarbonate. This is a particularly important item in the Windhoek waters, but owing to the deposition of calcium carbonate in the Pahl borehole pipes the analyses is not truly representative of the fissure water. As is to be expected in very hot waters containing alkali bicarbonate or carbonate the silica content is high. To maintain a balance between cations and

IONS	Windhoek Pahl spring borehole		Windhoek Jan Jonker boreholes		Gross Barmen		Outjo District (Position unknown)		Warmbad	
	mgm per litre	N /1,000	mgm. per litre	N /1,000	mgm. per litre	N /1,000	mgm. per litre	N /1,000	mgm. per litre	N /1,000
$NH_4^{\cdot}$	nil	—	nil	—	tr.	—	—	—	—	—
$Na^{\cdot}$	212.2	9.23	58.6	2.55	132.2	5.75	293.5	12.77	550.3	23.93
$K^{\cdot}$	101.3	2.59	76.4	1.95	134.2	3.43	155.4	3.98	9.6	0.25
$Mg^{\cdot\cdot}$	12.1	1.01	40.4	3.37	2.64	0.22	86.9	7.15	5.5	0.46
$Ca^{\cdot\cdot}$	17.9	0.89	90.1	4.50	37.9	1.89	69.2	3.46	163.6	8.13
$Al^{\cdot\cdot\cdot} + Fe^{\cdot\cdot}$	tr.	—	tr.	—	—	—	—	—	0.9	?
Sum of cations	343.5	13.72	265.5	12.37	306.94	11.29	605.0	27.36	729.9	32.82
Cl'	63.0	1.78	42.0	1.18	127.9	3.61	543.2	15.32	709.6	20.01
NO_3'	nil	—	nil	—	—	—	nil	—	tr.	—
SO_4''	176.4	3.67	141.6	2.95	373.8	7.78	411.4	8.57	546.6	11.38
HCO_3'	224.6	3.68	241.2	3.95	n.d.	—	211.1	3.46	857.7	1.41
SiO_3''	177.2	4.66	151.9	4.00	n.d.	—	—	—	—	—
Sum of anions	441.2	13.79	576.7	12.08	501.7	11.39	1,165.7	27.35	2,113.9	32.80
H_2SiO_3	—	—	—	—	n.d.	—	44.0 («silicates»)	—	109.2	—
Dissolved H_2S	present	—	present	—	4.25	—	n.d.	—	n.d.	—
Total sum of items determined ...	784.7	—	842.2	—	812.89	—	1,814.7	—	2,223.1	—
Items sought but not found	—		—		—		NO_2', PO_4''', $HBO^{8'}$		—	
Date and analyst ...	1926, Versuchs & Lehranstalt für Brauerei, Berlin				1911, C. Grimme		1929, G. Cock		1914, W. Versfeld	
Temperature of water ^0C.	79.8		33—35		65		46		37.5	

anions it has had to be calculated as SiO_3'' in the Windhoek and Gross Barmen waters.

Fluorine was not looked for, but judging by the incidence of dental fluorosis it is undoubtedly present.

Tests made in 1931 by Gevers[3] with a calibrated electroscope showed that the scalding waters of Windhoek are only slightly radioactive. The cooler water from the Jan Jonker and other boreholes is much more radioactive, the highest value obtained being 12.23 Maché units at the boreholes of the South West Breweries, Ltd.

VII. Spring Deposits

At Windhoek there has been extensive deposition of calcareous tufa, the layer at Gross Windhoek having a maximum thickness of 2 ½ to 3 m. Recently deposited material in the pipes at the Pahl borehole has the following composition :

$CaCO_3$	98.24
$CaSO_4$	0.41
$BaSO_4$	0.38
Fe_2O_3	0.80
MgO	tr*
Insoluble	tr
	99.83

Analyst : J. P. van der Westhuyzen.

VIII. Source of the Waters and their Heat

The Windhoek springs rise almost vertical fissures along which, in possibly Cretaceous times, there was volcanic activity. Although the fissures are of volcanic origin, Gevers[3] regarded the water as almost certainly meteoric. Rain falling on the mountains just south of the Windhoek valley and some 700 to 800 m. (2,296 to 2,624 ft.) higher is believed to get into the fissures by percolating along a bed or beds of permeable quartzites that are interleaved with garnet-biotite schists. These beds dip northwards towards the springs, first steeply and then at about 20° to 25°, and the uppermost quartzite would lie about 2,000 to 3,000 m. before Windhoek. No geothermal measurements have been made in the Territory, and Gevers assumed the geothermal step to be about 33 m. per 1°C. In Southern Rhodesia, however, measurements at the Globe and Phœnix Mine gave average values of 67.5 m. and 65.3 m. per 1°C. for magnesian schists and Archaean granite respectively.[7] Assuming the applicability of these figures to the rocks around Windhoek, the water would have to descend to about 4,000 m. (13, 124 ft.) to attain a temperature of 80°C. As the spring fissures are of volcanic origin, it may quite safely be assumed that they penetrate to a great depth, and it seems likely that even at fairly great depths they are sufficiently open to permit percolation of water.

There has been no post-Mesozoic volcancity in the Territory, and it is extremely probable that all the thermal springs are of meteoric origin, the water becoming heated as a result of the normal geothermal gradient supplemented, perhaps, by frictional heat developed by the slight movements that still take place along several of the spring faults.

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ANALYSES OF SOUTH AF

Locality	Natal Spa, Natal		Badplaats, Transvaal		Evangelina Transvaal		Tugela, Transvaa	
Formation	Archaean granite		Archaean granite		Archaean gneiss		Archaear gneiss	
Temperature of water °C	44.4		50.0		32.5		40-41	
Ions	mgm. per litre	N /1,000	mgm. per litre	N /1,000	mgm. per litre	N /1,000	mgm. per litre	N
$NH_4^{\cdot}$	0.05	0.003	nil.	—	n.d.	—	n.d.	
$Li^{\cdot}$	n.d.	—	n.d.	—	tr.	—	nil	
$Na^{\cdot}$	75.7	3.29	119.1	5.18	343.8	14.95	364.3	15
$K^{\cdot}$	n.d.	—	3.7	0.092	n.d.	—	6.6	0
$Mg^{\cdot\cdot}$	1.5	0.13	1.8	0.15	29.4	2.43	0.6	0
$Ca^{\cdot\cdot}$	5.7	0.29	8.0	0.40	78.8	3.94	123.7	6
$Al^{\cdot\cdot\cdot}$	2.5	(0.18)	n.d.	—	0.53	0.06	nil	
$Fe^{\cdot\cdot}$			n.d.	—	1.4	0.05	tr.	
$Mn^{\cdot\cdot}$	n.d.	—	nil	—	n.d.	—	n.d.	
Sum of cations	85.45	3.893	132.6	5.822	453.93	21.43	495.2	22
F'	n.d.	—	10.5	0.55	3.2	0.17	0.4	0
Cl'	46.0	1.30	134.9	3.80	443.0	12.49	418.0	11
Br'	n.d.	—	n.d.	—	n.d.	—	n.d.	
NO_3'	nil	—	nil	—	n.d.	—	n.d.	
SO_4''	23.4	0.49	24.0	0.50	215.9	4.49	503.9	10
PO_4'''	n.d.	—	n.d.	—	nl.	—	nil	
HCO_3'	—	—	33.6	0.55	191.3	3.14	27.0	0
CO_3	54.6†	1.82	18.0	0.60	—	—	—	
Sum of anions	124.0	3.61	221.0	6.00	853.4	20.29	949.3	22
SiO_2 (with N /1,000 as $SiO_3^{\cdot\cdot}$)	63.2	2.11	55.0	1.83	48.2	1.61	73.0	2
Dissolved H_2S	n.d.	—	0.21	—	nil	—	nil	
Total sum of items	272.65	—	408.81	—	1,355.53	—	1,517.5	
$NaHCO_3$	—	—	—	—	n.d.	—	n.d.	
Na_2CO_3	—	—	32	—	n.d.	—	n.d.	
pH	—	n.d.	8.95		7.15		7.85	
Sought but not found	NO_2'		NO_2'		Cu, SO_3'' B		Cu, B	
Date and Analyst	c. 1946, J. Gray		1948, W. Sunkel and P. Kok		1941, C. J. Liebenberg		1941,C.F.J. der Wal	

† Total Carbonates expressed as CO_3''.

* Includes 0.10 mgm. per litre of SH′ wit

RMAL WATERS

ysland, svaal aean de-gneiss 6	Sulphur Springs, Transvaal Pongola system 31		Machadodorp, Transvaal Pretoria series 27.—28.5		Die Oog, Transvaal Rooiberg series (felsites) 39.5		Riffontein, Transvaal Rooiberg series. 28.3—29.0	
N /1,000	mgm. per litre	N /1,000	mgm. per litre††	N /1,000	mgm. per litre††	N /1,000	mgm. per litre	N /1,000
—	n.d.	n.d.	—	n.d.	0.1	0.006	nil	—
—	nil	—	0.35	0.050	0.074	0.011	n.d.	—
4.29	26.5	1.15	54.7	2.38	46.4	2.02	235.0	10.22
0.15	n.d.	—	2.0	0.051	0.38	0.010	6.0	0.18
4.66	2.0	0.17	3.2	0.27	0.84	0.070	3.0	0.25
2.93	4.7	0.23	12.8	0.64	18.6	0.93	30.1	1.51
—	nil	0.043	1.5	(0.10)	n.d.	—	n.d.	—
—	1.2				n.d.	—	tr.	—
—	nil	—	n.d.	—	n.d.	—	nil	—
12.03	34.4	1.593	74.55	3.491	66.394	3.047	274.1	12.16
0.21	nil	—	n.d.	—	5.7	0.30	8.8	0.46
2.99	5.0	0.14	52.2	1.49	24.8	0.70	248.5	7.01
—	n.d.	—	n.d.	—	nil	—	n.d.	—
—	0.56	0.009	n.d.	—	0.01	0.0002	nil	—
0.35	7.0	0.15	5.3	0.11	8.2	0.17	38.4	0.80
—	nil	—	n.d.	—	nil	—	n.d.	—
8.30	—	—	54.9	0.90	109.8	1.80	237.9	3.90
—	39.0†	1.30	—	—	—	—	nil	—
11.85	51.56	1.599	112.5*	2.503	148.51	2.9702	533.6	12.17
2.87	44.0	1.47	26.7	0.89	42.0	1.40	48.2	1.61
—	n.d.	—	0.67	—	nil	—	nil	—
—	129.96	—	214.42	—	256.904	—	855.9	—
—	n.d.	—	n.d.	—	n.d.	—	202	—
—	—	—	—	—	—	—	nil	—
45	9.1		n.d.		n.d.		7.8	
	Ba$^{\cdot\cdot}$, Sr$^{\cdot\cdot}$, NO_2', B		—		Cr, As, Br$'$.		NO_2'	
,J. van Walt	1946, Messrs. McLachlan nad Lazar		1913, P.D. Hahn		1939, C.F.J. van der Walt		1948, W. Sunkel and P. Kok	

.003. †† Also determined — CuO 0.016, and B 0.08 mgm. per litre.

ANALYSES OF SOUTH AFRIC

Locality	Warmbaths, Transvaal		Buffelshoek Transvaal		Malmesbury, C.P.		L Tr
Formation	Bushveld granite		Bushveld granite (see p. 220)		Cape granite		Waterb
Temp. of water °C	51.9		30.6		32.9		3
Ions	mgm. per litre	N /1,000	mgm. per litre	N /1,000	mgm. per litre	N /1,000	mgm. per litre
NH_4	n.d.	—	n.d.	—	n.d.	—	nil
$Li^{\cdot}$	n.d.	—	nil	—	n.d.	—	nil
$Na^{\cdot}$	132.5	5.76	151.6	6.59	371.3	16.15	8.2
$K^{\cdot}$	2.9	0.07	5.7	0.15	tr.	—	2.8
$Mg^{\cdot\cdot}$	1.8	0.15	4.7	0.39	3.8	0.31	5.6
$Ca^{\cdot\cdot}$	13.0	0.65	27.1	1.35	51.6	2.58	26.4
$Ba^{\cdot\cdot}$	n.d.	—	n.d.	—	n.d.	—	nil
$Al^{\cdot\cdot\cdot}$	n.d.	—	nil	—	1.5	0.17	tr.
$Fe^{\cdot\cdot}$	n.d.	—	nil	—	0.77	0.028	tr.
$Mn^{\cdot\cdot}$	n.d.	—	n.d.	—	n.d.	—	n.d
Sum of cations	150.2	6.63	189.1	8.48	428.97	19.238	43.0
F'	11.0	0.58	6.6	0.35	n.d.	—	0.4
Cl'	85.2	2.40	138.5	3.91	561.1	15.81	3.6
Br'	n.d.	—	n.d.	—	0.91	0.011	n.d
NO_3'	nil	—	nil	—	n.d.	—	nil
SO_4''	12.1	0.25	35.1	0.73	52.2	1.09	tr.
PO_4'''	n.d.	—	n.d.	—	n.d.	—	n.d
HCO_3'	—	—	213.5	3.50	79.7	1.31	122.0
CO_3''	102.0*	3.40	nil	—	—	—	nil
Sum of anions	210.3	6.63	393.7	8.49	693.91	18.221	126.0
SiO_2 (with N /1,000 as SiO_3'')	48.2	1.61	44.8	1.49	56.0	1.86	18.9
Dissolved H_2S	nil	—	nil	—	7.44	—	nil
Total sum of items	408.7	—	627.6	—	1186.32	—	187.9
$NaHCO_3$	—	—	—	—	—	—	—
Na_2CO_3	137.8	—	93.3	—	—	—	15.9
pH	8.65		7.1		n.d.		7
Sought but not found	—		—		—		NO_2'
Date and Analyst	1940, W. Sunkel		1944, W.. Sunkel		1922, J. Muller		1941, C der W

* Total carbonates expressed as CO_3''.

Brandvlei, C.P.		Warmwaterberg, C.P.		Olifants valley, C.P.		Zwartkops, C.P. (borehole)	
Cape system		Table Mt. series		Table Mt. series		Cape system (see p. 220)	
64.2		45.6		50-51		54.5	
mgm. per litre	N/1,000	mgm. per litre	N/1,000	mgm. per litre	N/1,000	mgm. per litre	N/1,000
nil	—	nil	—	nil	—	nil	—
nil	—	n.d.	—	n.d.	—	n.d.	—
14.5	0.63	22.3	0.79	23.0	1.00	145.8	6.30
2.3	0.06	8.9	0.23	8.9	0.23	10.2	0.26
1.0	0.082	6.0	0.49	6.1	0.50	16.4	1.35
2.9	0.15	18.0	0.90	12.0	0.60	10.0	0.50
n.d.	—	n.d.	—	n.d.	—	n.d.	—
0.85	0.094	0.6	(0.04)	2.1	(0.15)	2.8	0.31
0.14	.005					3.2	0.11
n.d.	—	0.9	0.033	2.9	0.11	1.0	0.036
21.69	1.021	56.7	2.663	55.0	2.59	189.4	8.866
nil	—	0.20	0.015	0.30	0.016	0.50	0.036
16.5	0.47	37.3	1.05	37.6	1.06	237.9	6.71
nil	—	n.d.	—	n.d.	—	nil	—
nil	—	nil	—	nil	—	nil	—
2.8	0.060	10.6	0.22	12.0	0.25	33.6	0.70
n.d.	—	n.d.	—	n.d.	—	0.4	0.013
24.2	0.40	85.4	1.40	61.0	1.00	82.4	1.36
—	—	nil	—	nil	—	nil	—
43.5	0.93	133.5	2.685	110.9	2.326	354.8	8.819
29.5	0.982	15.0	0.50	31.0	1.12	18.0	0.60
nil	—	nil	—	nil	—	0.53	—
94.69	—	205.2	—	196.9	—	562.73	—
—	—	16.8	—	nil	—	nil	—
9.5	—	nil	—	nil	—	nil	—
n.d.		7.5		7.35		6.95	
NO_2'' I'		—		—		—	
946, J. Muller		1947, W. Sunkel and P. Kok		1947, W. Sunkel and P. Kok		1948, W. Sunkel and P· Kok	

ANALYSES OF SOUTH AFRICAN

Locality	J.R.I. nr. Odendaalsrus. O.F.S. (borehole)		T.G.I. nr. Trompsburg, O.F.S. (borehole)		Florisbad, O.F.S.		Stinkfontei C.P.	
Formation	Ventersdorp syst. (lavas)		Pre-Karroo norite		Ecca series		Beaufort se (Lower sta	
Temp. of water °C	34		37.2		28-30		28.7	
Ions	mgm. per litre	N /1,000	mgm. per litre	N /1,000	mgm. per litre	N/ 1,000	mgm. per litre	N /1
$NH_4^{\cdot}$	n.d.	—	nil	—	0.62	0.035	n.d.	—
$Li^{\cdot}$	n.d.	—	n.d.	—	0.69	0.099	n.d.	—
$Na^{\cdot}$	1246.0	54.20	2665.3	115.88	744.1	32.35	209	9
$K^{\cdot}$	9.0	0.23	16.1	0.41	nil	—	3	0
$Mh^{\cdot\cdot}$	7.3	0.60	20.0	1.65	0.79	0.065	15	1
$Ca^{\cdot\cdot}$	97.8	4.88	473.2	23.66	93.4	4.66	4	0
$Sr^{\cdot\cdot}$	n.d.	—	7.2	0.16	n.d.	—	n.d.	—
$Ba^{\cdot\cdot}$	n.d.	—	n.d.	—	0.74	0.011	n.d.	—
$Al^{\cdot\cdot\cdot}$	n.d.	—	n.d.	—	0.85	0.094	nil	—
$Fe^{\cdot\cdot}$	n.d.	—	n.d.	—	0.33	0.012	nil	—
Sum of cations	1360.1	59.91	3181.8	141.76	841.52	37.326	231	10
F'	1.0	0.053	3.6	0.19	n.d.	—	6.0	0
Cl'	2101.6	59.27	4245.8	119.74	1304.6	36.79	82	2
Br'	n.d.	—	nil	—	n.d.	—	n.d.	—
I'	n.d.	—	n.d.	—	tr.	—	n.d.	—
NO_3'	nil	—	nil	—	0.38	0.006	nil	—
SO_4''	13.5	0.28	946.2	19.70	2.3	0.047	163	3
PO'''	n.d.	—	n.d.	—	tr.	—	n.d.	—
HCO_3'	48.8	0.80	nil	—	18.7	0.31	268	4.
CO_3''	nil	—	60.0	2.00	—	—	12	0.
Sum of anions	2164.9	60.403	5255.6	141.63	1325.98	37.153	531	10
SiO_2 (with N /1,000 as SiO_3'')	11.0	0.36	26.0	0.86	21.9	0.74	28	0
Dissolved H_2S	—	—	—	—	tr.	—	16	—
Total sum of items	3536.0	—	8463.4	—	2189.4	—	806	—
$NaHCO_3$	nil	—	nil	—	—	—	252	—
Na_2CO_3	nil	—	nil	—	—	—	21	—
pH	8.25		8.95		n.d.		9.1	
Sought but not found	—		$Mn^{\cdot\cdot}$, NO_2''		Rb, Cs (sp.)		NO_2'	
Date and Analyst	1947, W. Sunkel and P. Kok		1948, W. Sunkel and P. Kok		1916, M. M. Rindl		1948, W. Sun	

* From a 1938 analysis.

Cradock, C.P. Beaufort series (Lower stage)		Fort Beaufort, C.P. Beaufort series Lower stage)		Aliwal North, C.P. Beaufort series (Upper stage)		Chipise, Transvaal Karroo system	
29.-31.3		27.0-29.0		36.9		65	
mgm. per litre	N/1,000	mgm. per litre	N/1,000	mgm. per litre	N/1,000	mgm. per litre	N/1,000
n.d.	—	n.d.	—	n.d.	—	n.d.	—
nil	—	n.d.	—	2.3	0.33	n.d.	—
49.9	2.17	163.7	7.12	344.5	14.98	140.9	6.13
0.6	0.015	2.2	0.051	6.2	0.16	2.2	0.06
0.6	0.049	3.0	0.25	5.5	0.46	1.3	0.11
3.0	0.150	11.0	0.55	83.8	4.19	6.0	0.30
n.d.	—	n.d.	—	n.d.	—	n.d.	—
n.d.	—	n.d.	—	n.d.	—	n.d.	—
nil	—	} 1.2	(0.09)	7.3	0.81	n.d.	—
nil	—			n.d.	—	1.4	0.05
54.1	2.384	181.1	8.061	449.6	20.93	151.8	6.65
5.8	0.30	13.2	0.69	n.d.	—	n.d.	—
24.9	0.70	189.9	5.36	605.3	17.05	139.1	3.92
n.d.	—	nil	—	1.6	0.02	n.d.	—
n.d.	—	n.d.	—	n.d.	—	n.d.	—
n.d.	—	nil	—	n.d.	—	tr.	—
16.8	0.35	36.0	0.75	68.6	1.43	27.6	0.57
n.d.	—	n.d.	—	n.d.	—	n.d.	—
9.8	0.16	39.7	0.65	29.4	0.48	103.5	1.70
27.0	0.90	16.8	0.56	44.2	1.47	12.0	0.40
84.3	2.41	295.6	8.01	749.1	20.45	282.2	6.59
40.2	1.34	37.0	1.23	14.8	0.49	66.7	2.22
2.3	—	6.2	—	0.79	—	1.6*	—
180.9	—	519.9	—	1214.19	—	502.3	—
—	—	84	—	—	—	—	—
—	—	30	—	—	—	—	—
9.6		9.0		n.d.		n.d.	
—		Mn..		—		—	
1946, W. Sunkel and P. Kok		1947, W. Sunkel and P. Kok		c.1913, P.D.Hahn		1944, H.S. Schoeman	

Rapports relatifs à la quatrième question :

CONDITIONS DES EAUX SOUTERRAINES DANS DIVERS PAYS

1) GROUND WATER STUDIES IN HYDERABAD (INDIA)

by

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Item I. Infiltration (absorption by ground); factors which determine the portion of rainfall which infiltrates while also considering the effects of plants, the configuration of the soil etc., and the importance of these influences.

Introduction.

The Hyderabad State comprising an area of about 84.000 square miles, lies between Lat. 15°-0'-0''; 21°-0'00-'' and Long. 74°-30-'0''; 81°-30'-0''. It has the following geological formations :—

ARYAN GROUP :	1. Alluvium	Recent
	2. Older alluvials of Godavery, Tungabhadra, Kistna and Bhima Rivers.	Pleistocene.
	3. Deccan Traps	Eocene, Oligocene.
	4. Gondwanas	Ranging from Jurassic to Carboniferous.
PURANA GROUP :	5. Vindhyans	Pre-cambrian
	6. Cuddapahs	
VEDIC GROUP :	7. Peninsular Gneisses	Archaeans
	8. Dharwars	

Of the above mentioned formations, the Archaeans composed of various granitoid gneisses and the Dharwars, occupy about half the area. The next extensive formation of Deccan Traps covers an area of 32.000 square miles. The remaining area of more or less 10.000 sq. miles, is composed mainly of various sedimentary series (i.e. Sandstone, Shales, Limestones, Quartzites etc.), with small patches of the older end newer alluvials fringing the rivers Godaveri, Kistna, Tungabhadra and Bhima.

The present physiography is the result of long continued denudation and earth movements, but the Deccan Trap configuration appers in comparison to be more undulating with abrupt rises and falls, and covered for the most part, particularly in the low lying and valley zones, with black cotton and loamy soils. The sedimentary formations are also likewise covered by similar type of soils, but the Gneissic area by both black and mooram soils. Plate I is a Geologic and Hydrographic map of the State.

The maximum average rainfall in the area is 42.6'' and the minimum average 17.57 inches. Areas receiving rainfalls of 25 inches and less (vide Plates I and II), are classed as Famine Zones. These drought contditions, and the inequities of even this low rainfall, results of the failure of even drinking water supplies. About one-third of the area lying to the West, South-west and South of the State falls under this category and as a consequence the villagers are put to much hardship, particularly during the four summer months.

Though the rainfall in the State does not appear to have appreciably diminished, still its economic value is none the less greatly impaired by ruthless destruction of forests and vegetation and by other artificial features. This has resulted in the lowering of the groundwater table, while the run-off into streams and rivers has greatly increased. The once existing pit like wells which supplied drinking water to the villages, having now gone dry, drinking water problem had become acute. The Government having foreseen the terrible consequences, started a scheme of providing hygenically constructed draw-wells at Government expense for drinking water supplies in all the villages in the State, the work being confined first to the Famine Zone. To this end, a Department was created in 1927. This Department is almost unique in the whole of India, in that for the first time well sinking was taken up as a comprehensive and extensive problem affecting the well being of the villagers. Being pioneers in the line, the Department had to evolve its own technique, make mistakes and learn several valuable lessons.

So far, more than 5.000 wells have been constructed ranging in depth from 30 to 115 feet maximum. Besides there are at present 2.000 wells under construction. Plate II gives the details of the area where wells have been completed, where they are in progress, where work has to be started and where survey is in progress. The remaining area left blank within the thick Famine Zone line, has still to be surveyed.

Of the many surface phenomena, rainfall is one which is most vital for the survival of man. Rainwater is disposed of in various ways, such as evaporation, run-off and percolation. The last factor, which is more or less synonymous with infiltration, is solely responsible for the formation of underground waters.

Percolation

Percolation is dependent on many factors, such as climatology, the intensity, duration and interval of rainfall, the nature, consistency, thickness and the disposition (inclination) of the soil cover; the type and closeness of vegetation and other local impediments and artidifcial features on the surface and finally on the geological and structural peculiarities underlying the soil cover.

Climatology

The effect of climatology on percolation is manifested in many ways. Evaporation is greatest on flat country and greater in hot weather than in cold, varies with the seasons and with the humidity of the atmosphere.

Intensity, Duration, and Interval Of Rainfall

Usually at the outbreak of the monsoons, absorption is much greater since the surface is dry and thirsty and the loss by evaporation is comparatively less than when the ground is either moist or saturated with rainwater. Generally speaking — other factors being normal — the greater the rainfall the greater is the percolation. Torrential rains of short duration , however, get lost as run-off. A continuous shower of rain after soaking into the ground to the extent that it can take up at a time the rest will be lost as run-off. Small showers distributed at intervals, thus have greater chances of percolation as the surface during the interval would have been rid of its extra water by soakage into the lower layers and be in a condition to absorb further supplies from the surface.

Nature, Consistency, Thickness, Disposition (Inclination) of Soil Cover

In general the soil cover performs one of the most useful functions in first becoming wet after rainfall, and assisting in the downward percolation of a part of that water into the zone of saturation. But the amount of percolation to this zone — which is discussed elsewhere — depends on the nature of the soil cover. A bare rocky surface devoid of any soil cover cannot absorb any water, except very minute quantities within the joints and pore spaces; but even such a surface if soil covered, will show material changes in its capacity for water. Of the many types of soils, black cotton soil, which is in a colloidal state of division, can absorb the greatest amount of water — about 52 percent — which is far greater than the quantity that can be held in a coarse or medium grained mooram (weathered gneiss) soil, as shown below by the porosity figures of different soils and decomposed materials:—

	Porosity	Gallons of water per cubic feet.
1. Soils	50%	3.00
2. Clays	45%	2.80
3. Sands & Gravels	30 to 40 %	1.8 to 2.5
4. Completely decomposed rocks containing much clayey matter	30 to 40 %	1.8 to 2.5
5. Highly decomposed rocks	16 to 24 %	1.0 to 1.5
6. Much weathered rocks	4 to 16 %	0.25 to 1.00
7. Slightly weathered rocks	1.5 to 3.0 %	0.10 to 0.18
8. Fresh Rocks	0.05 to 0.30 %	0.003 to 0.018

The figures given above indicate that a cover of soil has the greatest absorption. If the thickness of soil cover increases the water holding capacity will also increase in direct proportion. Also, the finer the soil the greater will be its porosity. The disposition (inclination) of the soil cover is another important factor in controlling the percolation. If the gradient is gentle or the ground is almost level, the percolation will be at its maximum, but the same is reduced by the increase in the slope of the ground.

A reference to Plate III will bring out the significance of these features. Out of many wells provided in Raichur District, a few have been taken and plotted according to their approximate Mean Sea Levels (M.S.Ls). Of the two wells « A » and « B » in village Bhogapur, the former has passed into greater thichness of weathered gneiss underlying the soil than the latter. Thus in well « A » the combined thickness of soil and the underlying decomposed soft material is almost double than in well « B ». Thus hte well « A » has a greater intensity of percolation and the yield is as much as 1 ½ times that of well « B ». In Kelarhatti village this is to some extent reversed, since the physiography round the well « A » is more favourable than that around B. Again at Kamalpur the combined thickness of soil and the underlying soft material in A is 15 feet and that in B is 10 feet; the former has greater percolation and therefore greater recuperation than the latter. Taking the case of Sindhnur Taluq, the two wells A and B at Sultanpur village show decided differences in their yield of water due to variable thickness of soil cover. In the former the soil extends to 15 feet and in the latter to 10 feet, and hence the recuperation is greater in A being 360 gallons per hour, than in B where it is only 250 gallons per hour; the 5 feet of extra soft material in the former which can obsorb a greater amount of water than the material in the latter at this level contributes to greater soakage and yield.

Vegetation, Forests etc.

The factors dealt with above are again modified according to the nature and closeness of vegetation. A grass-grown surface will naturally absorb a greater amount than a bare surface, since the plants arrest the immediate flow of water on the surface and effect greater percolation in the interval through the ramifications of their roots. Similarly forests also exert beneficial effects in increasing the economic value of the absorptive power of the soils. The part played by vegetation, specially forests is far more important than is generally reaslied. It may be relevent to point out at this stage the disastrous results brought by destruction of forests and vegetation, in the diminution of rainfall, lowering of water-table, quick run-off with the consequent erosion of soil . There are instancss even within living memory. Raichur District, which was once covered by thick forest with large number of tanks and plenty of water, has, due to ruthless destruction of forests and vegetation been now converted into a dry and dreary waterlees country, resulting in general famine and scarcity of even drinking water. The many existing ashler lined step-wells which were once sources of good drinking water, have either gone dry or become saline and bear testimony to the once prosperous nature of the country. Another indirect result of deforestation and the consequent lowering of water-table, has manifested itself by the appearance of salinity in soil and sub-soil making supplies of water in large areas unfit for consumption whether by human beings or by cattle. It is known that salinity, at least a part of it, is through the absence of surface and sub-soil drainage which would exist in areas with fairly good rainfall.

Other equally important phenomena brought about by the destruction of forests and vegetation may also be mentioned. The quick-run-off of rainwater through clearing of jungles and vegetation and burning away of even shrubs, has accelerated soil erosion, resulting in large tracts of cultivable land having been rendered useless for agricultural and other purposes. If this remains un-checked, it will denude, in course of time, almost all the soil on the surface, rendering the area barren thus resulting in famine and pestilence. As all the soil is thus being lost (since it is not formed as quickly from the decomposition of the rocks) the absorption will also gradually diminish, thus lowering the groundwater and eventually drying up the zone. The major portion of Raichur District (lying between the two rivers Tungabhadra and Kistna) is covered by black cotton soil and due to the peculiar configuration most of the soil lying in the Highlands has been washed away leaving bare rocks at the surface. These areas were once covered by fair thickness of soils yielding substantial returns by agriculture. Similarly the Deccan Trap area due to its abrupt rises and falls is also undergoing similar deterioration and it has been noted that usually the highlands are almost either bare of soil or are covered by only a very thin mantle of soil. The absence of soil cover has therefore, not only reduced the economic value, but has interfered with the ground-water resources due to which there is scarcity of drinking water supplies from such sources. No country therefore, can look with equanimity at this state of affairs brought about by thoughtless land use.

Forests and vegetation controll to a large extent the evaporation of water from soils, thus increasing the holding power in them. The relative evaporative properties of different surfaces, taking unity as evaporation factor for bare ground, are :

1. Pasture land	0.7
2. Cultivated land	0.8
3. Light forests and	0.6
4. Dense forests	from 0.2 to 0.4

The beneficial effects of forests and vegetation have however, been sometimes questioned, in view of the two opposing factors

1. Loss by transpiration, and
2. Gain by « intake » into the ground.
(On account of the interception of rainwater and formation of greater permeable material caused by the desintegrative action of the roots and prevention of evaporation).

Shrubs and plants which send down their adventicious roots not far down the surface absorb only the soil water after they have performed the important and useful function of greater soakage into the underlying layers. On the other hand huge trees, which send in their roots far down even to depths of groundwater table, may to a certain extent interfere with the resource, which may be felt only at places of very low rainfall and where the groundwater supply is highly localised and limited. With all these it may be concluded that forests exercise beneficial affects as regards water supply and stream flow and this though at present may not admit of regional consideration for want of sufficient data, may be regarded as applicable to large areas and in particular to certain localities.

The soil water therefore, after gradual percolation into the deeper layers, replenishes the store of permanent water. But how much of this soil water is available for absorption by the underlying layers is a point which has to be seriously considered. The black soil though highly porous, does not part with its water due to fineness of the material where the water is held by capillarity and molecular attraction. Hence, generally it is difficult to tap good supplies of ground-water in areas covered with such soil. In contrast, a loamy or sandy soil though it absorps a lesser amount of water, parts with most of its water into the underlying materials or into wells sunk in them. This shows that porosity is not the only criterion in the consideration of sub-soil yield, but the condition under which this water is held, and the ease with which soil can part with most of its water. That which is yielded is the « specific yield » and that retained is the « specific » retention; the two together forming the porosity of the materials.

Nature of Sub-soil

The nature of sub-soil is another factor which plays an important part in the percolation of water to the underlying decomposed material or rocks. In regions of black soil cover, due to the restricted subsurface flow and other concomitent chemical interchanges by the formation of alkali carbonates and sulphates, a sort of an impervious sub-soil layer is formed which arrests the downward percolation of water. Under such cases the underground water supply will not only be highly limited, but its quality will be such as to make it unfit for drinking or even irrigation purposes. If stagnation of water occurs over this impervious layer, deleterious salts from that layer may be gradually brought up to the surface during summer by capillarity fouling the soil and impairing the productivity of the same. Without going into the details of this subject and reverting to the matter under discussion, the next point arises as to the condition under which the permanent water is formed and the depths to which it extends.

For a source of good supply of groundwater, the materials from which it is to be obtained should also be considered. For instance, a soft underlying material of granular to gritty nature, is not only highly porous, but will part with its water more readily, there being a freedom to flow of the groundwater; in jointed rocks the supply may not be adequate as the water that is tapped in them is the water that is stored in the joints, crevices or bedding planes etc. If the joints are interconnected with extensive ramifications, such rocks would yield fair supplies of water, otherwise, the supply will be very limited. It is this feature that brings about anomalies in wells situated in such areas; one well tapping good supplies and another giving a poor yield. Wells or infiltration galleries to tap this source of groundwater should be located far below permanent watertable for the following obvious reasons:

Generally, above the zone of saturation there is a varied thickness ranging usually from 10 to 15 feet of another zone known as of « intermittent saturation » This zone of intermittent saturation is a very important medium through which the underground waters are replenished year by year. It is therefore, obvious that this zone cannot denude itself of all its water except perhaps the retentive water. Similarly the soil overlying this zone cannot also denude itself of all its water, though it may be dust dry at the surface. If conditions are not so, then the porosity of the soil and of the underlying decomposed rocks extending down to depths of more or less 40 ft. to which depths the wells in this State have to be normally excavated to reach the water bearing materials, would be so high, that the total annual rainfall will not be sufficient even to partly wet these materials, and no water can ever pass into the underlying zone of intermittent saturation to replenish the store of permanent water, nor will there be any universal rise in the water level in wells, which is a common phenomenon occuring all over during and after the rains. In other words there will be no source of groundwater at all if, during the monsoons the soil, sub-soil and the decomposed rocks down to depths of permanent water-table have only to be wetted barely even to the extent of their specific retention. A simple calculation based on the relative porosities of different materials would clarify the position. Taking a very conservative estimate of the soil cover in an area capable of absorbing and retaining about 30 percent of its volume, and leaving out the effects of evaporation etc., the amount of water contained in the average five feet depth of soil considered, would work out to 18 inches. The underlying decomposed materials, may likely also absorb a similar amount, but taking a low average figure of 20 percent for a thickness of the remaining 35 feet, to touch the permanent water-table, the amount of water absorbed would be 84 inches. In all, for this depth of 40 feet the amount of water held will be 18 + 84 = 102 inches. This amount would far exceed the annual rainfall in most areas and therefore it becomes obvious that a great proportion of the moisture in soils, sub-soils and the underlying decomposed rocks cannot be removed during the dry season. Even if we assume that only 20 percent of this moisture (i.e. 102) is removed, then that much of the amount has got to be replenished by the rainfall, which again means the soakage of either the entire annual rainfall or even more than that. Such a thing cannot happen since the rainfall is disposed off in various ways, such as evaporation, run-off and percolation. The last factor (percolation) is the one which replenishes the groundwater year by year, and therefore the entire rainfall cannot go this way. But under some exceptional conditions it may happen that more than 50 percent of the rain may soak into the ground and that a greater or perhaps the entire portion of it may go down to permanent water-level is evidenced by rises in water level even after a comparatively small part of the rain has fallen. True, there may be a rise in the low lying tracts which receive the run-off from the higher reaches of a large area which together with the local rainfall may very largely exceed the total rainfall, but this cannot apply to every situation. No doubt, evaporation, barometric variations due to oscillations of pressure on the surface, transpiration of plants and capillarity all play their respective part in depriving the soil of its water content, but with all that the total reduction of moisture cannot exceed a certain limit. It may be possible to make an estimate of moisture content in soils and in the zone of intermittent saturation by taking soil samples at various points and various depths from the surface, above saturation level at the end of dry season, but these are beset with so many difficulties and uncertainities, that after all, the results obtained may represent only approximations. The only course left to judge the groundwater resources and its annual replenishment is from the existing surface and structural features and the climatic conditions in the area.

The above discussion makes easy to explain the failure in many wells and infiltration galleries, the former for purposes of house or village water supplies and the latter for that of towns. Most of the wells that are sunk by the villagers or the infiltration galleries put down in hard materials are not usually taken down to permanent water level, and hence during summer they dry up, the latter perhaps sooner as the draw off is heavy. The supply dwindles and finally fails due to the hard nature of the jointed rocks; or even if situated in softer materials with freedom to flow of groundwater, because the bed is of limited thickness. Assuming for purpose of an illustration the permanent water in a certain area to be somewhere between 35 to 40 feet, the wells therein, which are generally shallow going not more than 30 to 35 feet deep, would not only prove seasonal wells, but will dry up in summer for the following reasons. The zone of intermittent saturation may be considered to be about 10 feet thick, overlying the permanent watertable, and during wet weather the wells will experience therefore, a rise of 10 feet in their water-levels. No doubt as the water is withdranw, the percolation will increase since the water-level falls below the surrounding water, but then the maximum head of water causing this percolation will be limited only to 10 feet. As the dry weather approaches, the water-level falls diminishing the percoilation into the wells in proportion to the fall, until finally the water-level falls to its lowest level, when the wells or infiltration galleries will go dry. Even in those cases where wells or infiltration galleries have penetrated only a few feet below the permanent water-level, the rate of flow may fall off considerably as the water-level recedes due to the diminution of the hydraulic pressure.

Again in cases of infiltration galleries in particular which are meant for large quantities of water to towns, the case requires the deepest and at the same time careful investigations of the area where it is to be located. Galleries in hard rock would hardly answer the needs, especially in summer when the zone of intermittent saturation gets almost denuded of its specific yield of water and the draw off is then restricted to the permanent water zone. The supply then being limited (far more than in the zone of intermittent saturation) due to water being stored in finer joints and smaller cracks of limited extent, eventually fails due to overdraw. In cases of granular (to gritty) materials already mentioned, well below permanent water zone, a series of trial bores in directions at right angles to each other may be made to find out the thickness and lateral extant of the material so as to calculate if water that can be yielded by the strata would be suffifient or not to cope with the demand particularly during the four summer months. Only in this way successful infiltration galleries can be located to meet the demands of large supplies. Since such a feature may not be available at every place or near every town, other methods wherein a permanent water zone of saturation can be substituted may be taken advantage of; this feature being available under beds of perennial streams and rivers.

A specific example where large sums of money were wasted with no advantage can well be cited here. When the existing water supply to a big town in Hyderabad State failed 4 times within twenty seven years of its completion, various other schemes were formulated to augment the water-supply. Eventually a scheme of laying infiltration gallery was proposed in a minor valley. No investigation of any importance was conducted except a cursory test of recuperation in a pit excavated in the bed of the valley to a depth of about six feet. On the merits of this test it was forecast that the scheme would come out entirely successful. The excavation of the gallery was started during summer 1941, and on reaching a depth of 20 feet, it was found that the rate of percolation from the sub-surface was not even one-twentieth of that observed in the experimental pit above mentioned. Evidently the material was not of the same consistency throughout and a mere pit made for testing the recuperation at one spot could not hold good and led to misleading conclusions and erroneous forecasts. In order to get even a near approximation a series of trial pits should be put down at regular intervals and their water yields tested. Even this cannot be taken as infailliable for reasons of varying physical structure of the material or the rock component. Hence it becomes possible that within a specified time even such an infiltration gallery will not be in a position to yield the required heavy demand, though when the time range is greater it may in all likelihood furnish the full supply. Assuming the porosity of the material in which the infiltration gallery was excavated to be as low as five percent and computing an area of 35 square miles of the valley zone above the site and a depth of permanent saturated material to be 10 feet (since the gallery was expected to be excavated to 20 feet and presumed to be within the zone of permanent water level) the amount of water that can be stored in such a mass would be 5.280 × 5.280 × 35 × 10 × 5/100 × 6.25 = 3.059.190,000 gallons. If only 40 percent is considered to be the yield, it would mean 1,223,680,000 gallons. This amount of water in the area under consideration having an average annual rainfall of 26 inches represents 25 percent of the total rainfall as percolation underground. This available water i.e. 1,223,680,000 gallons would certainly appear an enormous supply even for a fairly big town for a number of years; but unfortunately various underground structural peculiarities, the slope of the underlying material, the ease of flow groundwater or the intercommunication of the joints if the material is of a jointed nature and the impediments to the flow etc., so complicate the matters that it cannot be considered similar in every respect to the availability of surface water of the same magnitude. In contrast it may be pointed out that in alluvial deposits of very great extent and of the type met with in the Indo-Gangetic plains, composed entirely of sands with few layers of silt, copious and inexhausible supplies of water can be tapped even from small tube wells as the rate of flow is so rapid that the water-level in the bore cannot be lowered appreciably by heavy draw off.

Surface Relief

The relief of the land has a very important bearing on groundwater conditions Somewhere and at some time or the other the groundwater emerges on the surface. as seeps and springs. In a low lying flat country, composed of porous formations, the water very nearly fills up the rock and may be available in wells sunk to shallow depths. In a hilly or mountainous region the water that percolates into the ground will very soon tend to seep laterally and begin to flow as springs at different levels. In areas of highly undulating surface with ridges and knobs here and there, the percolation of water would also gradually seep into minor valleys and eventually into the major ones and emerge out as springs at some favourable spots. Under these circumstances the movement of groundwater in the resultant direction (one due to gravity downwards and another lateral due to head or pressure) will be comparatively high due to local declivities, sudden falls etc., of the surface.

It is therefore, clear that except in places where gullies or valleys have been eroded down to a water bearing stratum of where water emerges out of the ground in the form of spring, at all other places where the groundwater can be tapped by means of wells, the choice of site is of paramount importance.

How physiography and soil cover are of importance may well be illustrated by examples.

In a village on Traps at Osmanabad Taluq (Lat. 18°-10′ Long. 76°-6′) two well sites were chosen, one on a little rise with the surrounding ground all barren with outcrops of jointed rocks, another about 120 feet away after an abrupt fall of only 5 feet towards the slope of the ground and surrounded by agricultural land. The first well went down to 93 feet with barely 6 inches of recuperation whereas the second well at 60 ft. gave 10 feet overnight recuperation. This was certainly due to greater infiltration of water round the second well than in the first and also due to greater catchment area falling in line with the valley zone at the second well site; a very great difference in results even within short distances.

Of the other structural features controlling the water supply,some of the important ones may be enumerated. In a gneissic area, the existence of dykes or reefs markedly control the distribution of groundwater. In a sloping ground with a dyke traversing at right angles to the slope, a well to the rise of the dyke proved successful, yielding copious supply of water, whereas another, a few feet from the dyke towards the fall, proved dust dry. This is due to the dyke acting as a sub-surface dam in effectively barring the percolation of water to the other side of it. If the site on the fall side had been chosen farther away from the dyke to command sufficient catchment, then it would have also yielded enough water. A number of such cases came to the notice of the writer when engaged upon the scheme of sinking some 5.000 wells.

At Khanapur village (Lat. 16°-42′ Long. 77°-1′ 5″) Shahpur Taluq, Gulbarga District in the gneissic area, two well sites were chosen ,one nearer to a quartz reef which formed part of an irrigation tank bund, and another a little farther away from the same reef. The first well nearer the quartz reef had gone deeper than the well away from it, but did not yield as much recuperation as the latter notwithstanding its close proximity to the storage tank. This is due to the presence of quartz reef nearer the first well which in addition to reducing the catchment round the well effectively barred the flow of groundwater from the higher reaches above it. The second well, being far away, had comparatively a greater catchment which contributed to greater yield. This is a case where the quartz reef is almost in a vertical position, but in cases of inclined reefs also there have been marked differences in yields. In Gundhalli village (Lat. 16°-42′-40″, Long. 76°-59′-48″) Shahpur Taluq, Gulbarga District two wells A and B situated only a few feet apart gave quite surprising results in their yield of water. On examination it was found that well A met with a thin quartz reef at 15 feet from ground -level and having pierced it had gone into the underlying rock. The well B having gone 32 feet from ground-level met the same reef at 31 feet having touched the compact rock below that depth at 32 feet. In well A the recuperation was very poor, whereas in well B there was ample supply. This anomaly, curious as it may appear from surface features, is easily explained by the presence of the inclined quartz reef which acts as an impervious bar and retains above it all the sub-surface water, directing a greater amount along its fall into well B than into well A; the rock pierced below the reef in the two wells being quite compact and barren of water. All the draw of is directed therefore, towards B whereas A experiences a drain.

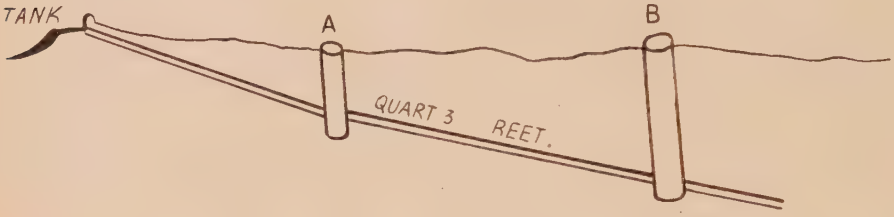

Plate IV gives a section from Madgota village (on the Krishna River) to Salgundi showing the relation of permanent water-table with surface configuration, observed in the year 1930. Along-side is given a skeleton map of parts of Raichur District to illustrate the section. The area is composed of the Archaeon Complex (i.e. Dharwars and Peninsular Gneisses). The vertical scale is exaggerated to bring out the significance of configuration in a more pronounced form. It will be interesting to note that (the highest point being between villages Chincherki and Topaldoddi with the ground sloping steeply on either side to North and South) the permanent water-level is nearer ground-level from Madgota to Markamdinni villages than between the latter village and Salgundi village although the country is comparatively less steep in the latter portion. This is due to a number of dykes cutting the area from East to West, North-east to South-west and North-west to South-east between these two villages, whereas between Markamdinni and Salgundi villages the area is almost free of such dykes. A few of the dykes traversing the area are shown both in the plan and section. Hence in the former case though the ground is sloping steeply the ground water is not easily allowed to seep to lower levels, the dykes therein acting as many impervious sub-surface dams in the area, whereas that between Markamdinni and Salgundi villages (being almost free from such dykes) the ground water has freedom to move laterally resulting in a lowering of the ground water table.

The schistose rocks i.e. Dharwars, though they form good material for the percolation of rainwater underground, do not yield that water at shallow depths due to the highly fissile (schistose) nature of the rocks which form planes of least resistence through which the percolating waters rapidly work their way to great depths. Wells in such material therefore — unless they are in small pockets within the gneisses extending to shallow depths — have to go deeper with prospects of only limited supplies. As an example at Tintini village (Lat. 16°-22′-0″, Long. 76°-40′-30″) Shorapur Taluq, Gulbarga district, a well, excavated in granitoid gneisses gave very good recuperation at comparatively shallow depth, whereas a few hundred feet away, another well excavated to nearly a depth of 60 feet in schistose rocks was almost a failure.

A similar instance was met with at Bijaswar village (Lat. 16°-33′-30″,Long. 76°-49′-15″) Shorapur Taluq, Gulbarga District, where the well in schistose material ever after going to 65 feet did not yield the expected recuperation, whereas that located on the gneissic area gave ample supply at shallower depth. Hence in some general cases wells in hard rock with good joints and cleavages yield better supplies since as the water-table is reached, the joints and cleavages become scarce and wide apart and almost disappear acting as an impervious layer over which the groundwater is stored and made available in wells.

Sandstones

Uncompacted sandstones being soft and porous retain and yield good supplies of water and generally therefore, wells in such material present no difficulties.

Limestones

The limestones being hard, fine and compact bedded rocks are generally unsuited as medium for percolation, but these are usually traversed at the surface by joints when some water no doubt percolates in them. Hence no large supplies can be anticipated in these materials — though adequate amounts can be obtained for village drinking water, provided the wells are located under good physiography and soil cover. No doubt in places of heavy rainfall « caverns » may occur formed by the solvent action of the percolating water, which if tapped would yield good supplies of water. In arid regions, however, they are scarcely met with.

Shales

Shales are most unsuited for percolation on account of their very low porosity. These are also however, traversed by joints and cracks through which surface water percolates downwards and hence some water may be obtained from them. Usually shales show poor recuperation as depth increases. The depth to which the wells have to go depends upon the depth to which the joints and crevices extend. In some places they may close up within 30 to 40 feet,in other places they may extend to 60 or even 100 feet. Under these conditions the wells have to be excavated to these depths to tap some dependable supplies of water. In Shahbad (Lat. 17°-30′, Long. 76°-55′), wells are generally 60 to 90 feet deep and even at that depth the supply is undependable.

The Table below gives the average total depth and water-level in wells in the Taluqs with the respective geological formations against each:—

Taluq.	Average depth of wells	Average depth of water in wells	Geological formations
Andola	52 ft.	22 ft.	Deccan Traps and Bhima Series of Sedimentaries
Shorapur	42 ft.	12 ft.	Gneissic
Shahpur	40 ft.	14 ft.	Gneissic and partly Bhima Series.
Sindhnur	33 ft.	10 ft.	Gneissic and Dharwar Patches
Kushtagi	37 ft.	10 ft.	Gneisses and Dharwars
Gangawati	32 ft.	8 ft.	— do —

It may be mentioned here than in Andola Taluq the Bhima Series of sedimentaries present faulted and disturbed zones to the south of Jewargi through Chennur, Gudur, Andola, Maradgi, Warchanhalli, Mudbal Buzurg and Mudbal Khurd to Madarki. In this zone therefore the greatest difficulty has been encountered in providing wells.

To the east of this the limestones run without much structural disturbance from Narboli to Sirwal.

The Deccan Traps to the rest of Shahpur — Jewargi Road is comparatively free from disturbance but the physiography presents some special features which have an intimate effect on the distribution of sub-surface water. The annual rainfall in this area is only 20 inches and with the country dissected by innumerable nalas, most of the water is drained into the Bhima River. To the natural adverse condition of the distribution of sub-surface water, may be added an artificial cause of considerable magnitude, namely the depletion of the already meagre resources by the opening up of the area by wells in villages. This entailed heavy and constant pumping during the course of excavation of wells and at a conservative estimate about 4.000 gallons from each of the wells have been thrown out daily for about 4 months. Hence during excavation the recuperation had gone on dwindling due to the cone of percolation getting exhausted and dry. It is not at all unfair to assume that when once the interstices in the underlying materials get resaturated during the succeeding years of rainfall the recuperation will no doubt improve considerably which can stand the strain of even severe drought. There have been many such cases of apprehension, of wells showing such slender recuperation during the course of excavation, but all these have very considerably improved in their yields and have withstood severe droughts.

Under the conditions explained above, it would be futile to expect unlimited supply of water within practicalbe depths. Some of the wells have gone down to about 90 ft. and the average depth in the shale area is easily about 70 ft. from ground-level. In arriving at the safe supply for a village even in these trying conditions, care has been taken to assure that adequate recuperation exists to meet the actual demands, allowing 20 gallons per head as the normal requirements, a figure which is much above the consumption.

Deccan Traps

Without going into the details of the history of this formation,it may be remarked that water finding in traps is beset with many difficulties, but if once the area is examined and the principles understood, the presence of ground-water can more or less be located with fair accuracy.

The influence of physiography has a very great bearing on the distribution of groundwater, though the nature and depth of soil cover are also equal or more important contributory causes.

The rocks of this formation usually show different physical characters, ranging from a few feet to more or less 100 feet, with varying ticknesses and extent of decomposed materials. Usually the highly altered types of mooram-like (1) materials, when situated at varying depths from the surface form good aquifers provided they have been in some way or the other saturated with water.

Exfoliating Rocks

The exfoliating rocks also yield good supplies of water through their highly weathered matrix, and these rocks are second in importance in relation to water yielding capacity, the first being the inter-bedded mooram beds (fully weathered rock beds).

Jointed Rocks

The highly jointed trap rocks lead the percolating waters far deep even to depths of 100 feet or more, when these rocks extend to this depth and hence wells in such rocks have not only to go deeper, but the yield is poor. If however, joints close down above those depths, fair supplies of water are also obtained in them.

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(1) The term « mooram » is here loosely applied to a. much decomposed rock which does not show its original characters. These range into pink, red,ashy to green shades, frequently containing the secondary zeolites.

As the beds of jointed rocks are more or less horizontally disposed a study of their sequence if available from some natural section or scarp or even from some artificial cutting or from existing old wells would furnish an insight into the existence of ground-water. This study warrants our assuming that the nature and sequence of beds so noted continue into the plateau itself. The different layers so observed if connected to Mean Sea levels (M.S.Ls) would form very valuable data on which to base our calculations both in regard to depths of wells and the yield of water. A large number of well sections taken in this formation have brought out clearly that under favourable conditions the zeolitic mooram-like layers form good aquifers yielding good supplies of water. However, these beds when situated under adverse physiography may not likely be saturated with water; therefore the same bed which is aquiferous in one place may be dust dry in another place. If such beds are not available within reasonable depths at any place,the next best aquiferous layer would be the exfoliating type; but in the absence of these two the jointed type may have to be depended upon for what it can yield which will however be governed by the local topographic conditions and other features mentioned above.

Much work has been done on this formation and so it is not possible to quote every individual case here, except a few ,with some generalised remarks as gathered from observations and experience. How physiography in this formation plays a very important part may here be exemplified.

Broadcastinghouse Aurangabad

A well at the Broadcasting House at Aurangabad (Lat. 19°-52′, Long. 75°-20′) was sunk to a depth of 64 feet in weathered rocks (soft material) with no yield of water. Further money was therefore invested on this well and a bore-hole was put down from its bed, piercing an additional depth of 40 feet. Even then the yield did not improve in any way. The reason for the failure was the adverse physiography being above a well defined ridge of 1.900 contour, which functioned as a water shed. Though there were thick hands of soft rock which could absorb good quantities of water, still they were practically dry, as no rainwater found its way to percolate into them, since most of it was lost as run-off into the valleys below. Knowing this feature fully well, there was really no object in putting a bore from the bed of the well to get water. It would have been perhaps reasonable if the well had been sunk in the valley zone and a bore put down from its bed for water to shoot out from the bore, but even here, unless structural features and physiography were carefully studied and understood, such a venture cannot be undertaken, and it therefore becomes all the more surprising to expect a sub-artesian or artesian phenomenon through a bore put down on a well defined knob forming the highest elevation in the area. The mouth level of the well was 1.931 feet above M.S.L. This will show from a comparison of the sequences given below, that though softer rocks are available for good depths, still due to adverse physiography no water was available in them.

State Railway Hotel, Aurangabad

The physiography round the State Railway Hotel, Aurangabed, though not so bad as the above, is similar from local consideration. The authorities were advised that bore-holes would not be successful at the place yet one was put down to a depth of 110 feet and another to 80 feet without the least indication of success. The rocks to this depth were as dry as dust. The mouth of the first bore was 1.868 feet and that of the second a few feet away 1.865 feet above M.S.L. Again by a comparison of the sequence given below it will be seen that though the geology is favourable, the adverse physiography has interfered with the water holding and yielding capacity of the rocks:—

1,950 to 1,913	Highly weathered, but vesicular at places.
1,913 to 1,885	Highly weathered exfoliating with harder patches.
1,885 to 1,880	Vesicular but weathered at places.
1,880 to 1,860	Weathered exfoliating rocks.

1,860 to 1,833	Exfoliating, weathered at places.
1,833 to 1,825	Harder rock but locally weathered.
1,825 to 1,806	Hard compact to massive rock.

All the rocks above 1,833 feet M.S.L. are soft which can store and yield water provided they are fed. Due to their relative position on the ridge, they had no such chance and the bores proved utter failures.

In Plate V are plotted the wells in position according to their approximate mean sea level along-side the natural section af the area. Excepting a few feet of soil cover, all the wells have more or less passed into similar materials in their respective horizons as that shown in natural section. The last two wells, Banwadi Military and Pyrasingh Mansingh at the corresponding level of the red friable mooram occurring in the natural section, have touched decidedly softer though not exactly the same material which is already mentioned in the report varies in its consistency from place to place according to the intensity of weathering. It is this layer that is responsible for the yield of water in the two wells. Other wells which are not mentioned in the paper but which are however plotted in the plate behave in a manner determined by these features.

A natural section from the scraps to the west of Osmanabad Town reveal the following nature of rocks:—

2,214 to 2,197	Weathered jointed rock.
2,197 to 2,172	(2,160) Pink to Reddish Mooram.
2,172 to 2,164	Much weathered jointed rock.
2,164 to 2,132	Compact jointed rock.
2,132 to 2,118	Weathered jointed rock.
2,118 to 2,090	Much weathered with exfoliating boulders.
2,090 2,109 } to 2,075	(2,070) Mooram Friable at top.
2,075 to 2,045	Comparatively weathered at places compact and bedded.
2,045 to 2,029	Compact rock but jointed at places.
2,029 to 2,018	Compact rock.
2,018 to 1,972	(1,962) Mooram like material harder towards bottom.
1,972 to 1,962	Weathered jointed, in parts exfoliating.

As may be seen from the above, it is obvious that softer beds and layers of mooramlike nature lie interbedded between rocks and provided such beds receive sufficient water by percolation through the overlying rocks they form good aquifers. It can also be observed that such thick inter-bedded bands of mooramlike material (weathered), persist to some extent laterally and at places disappear. From the section and the levels given in brackets, it was observed that these beds vary in their thickness from place to place as these are the remnants of denudation after weathering. This feature brings in the anomalies in wells situated close by in any one particular locality.

The perennial wells at Osmanabad are due to the existence of soft beds of mooram-like nature between levels 2,090 to 2,075 feet M.S.Ls.

The Police Line Well at Osmanabad, whose mouth is about 2,145 feet above M.S.L. has been sunk after advise to a total depth of 70 feet i.e. 2,075 feet M.S.L. when it resulted in yielding sufficient supplies of water ,though above this depth there was scarcely a few pots available. This is again due to the well having pierced the mooram-like layer occurring between the levels, but which at the place has thinned to a bed of 8 feet depth.

In Plate VI is also given logs of a number of wells with their water-levels in relation to ground-level in Deccan Trap area. The traverse (as shown in the skeleton map alongside) was conducted from North to South during the summer months when the water-levels in wells were at their minimum. These sections show more or less similarity in materials in their corresponding mean sea level with marked oscillations here and there. The relation of water-level to ground-level is also brought forth. From these it will be seen that for a good source of ground water supply all the features discussed above must obtain in a place and with any one of these factors proving adverse the resultant effect would also be adverse. Under these conditions it is not possible to arrive at any experimental data for any one factor or factors since the conditions governing the source of ground water derived by percolation of the surface waters are so varied as clearly brought out in the sections and the foregoing discussions.

Conclusion

The distribution of sub-surface water being a complex function controlled by several factors of which geological feature is one, others equally or even more important such as the rainfall and physiography have also to be borne in mind. The latter including the surface configuration, drainage, catchment and soil cover all play an important part. In a region with good rainfall one can look for nearly saturated condition of an aquifer, but in an arid country with hardly 20 inches of annual rainfall, only certain protected valleys with some catchment can alone benefit; the areas lying towards the highlands holding little or no water.

In addition to the surface and sub-surface conditions the disposition of natural growth of vegetation particularly big broad leaved trees should also be studied. These in general, show a directional trend which happen to be lines of disposition of good subsurface water, often corresponding to the valley or low lying parts in the area.

PLATE I.

GEOLOGICAL & HYDROGRAPHIC MAP OF HYDERABAD

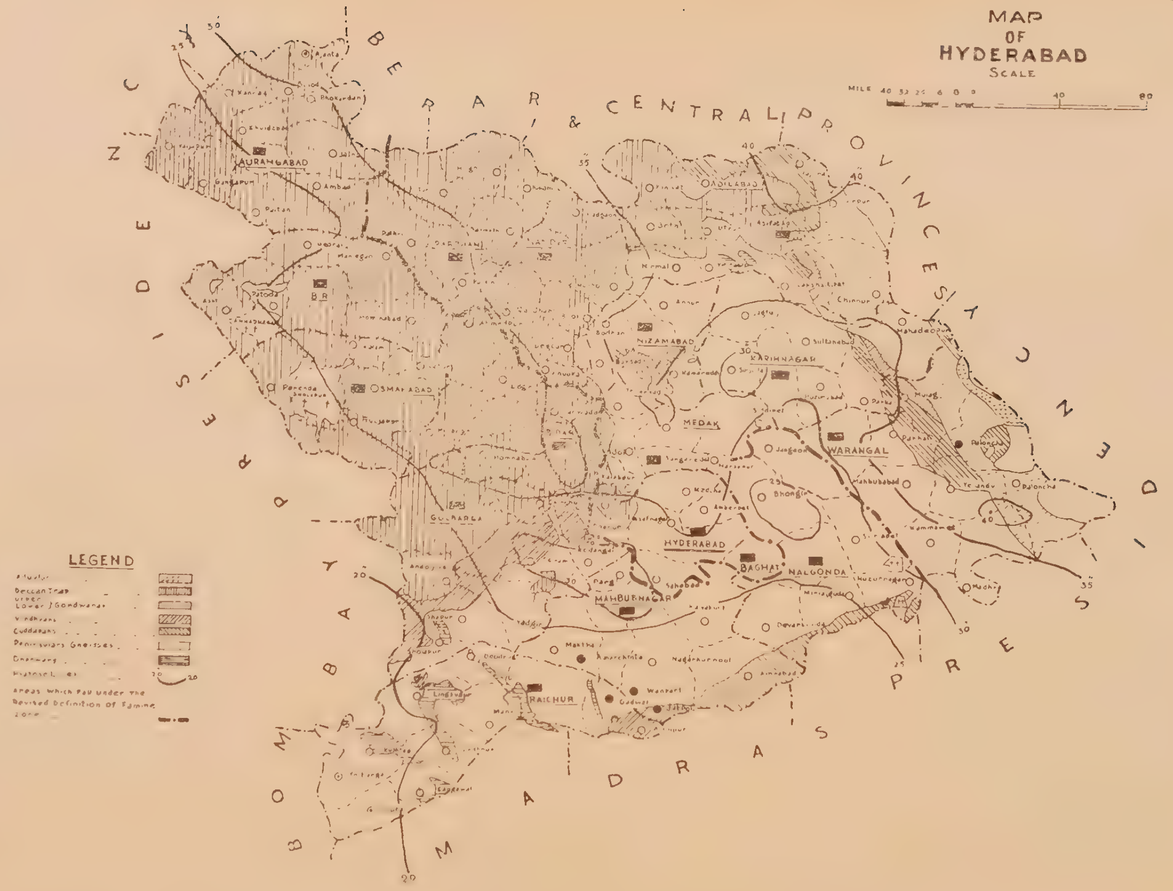


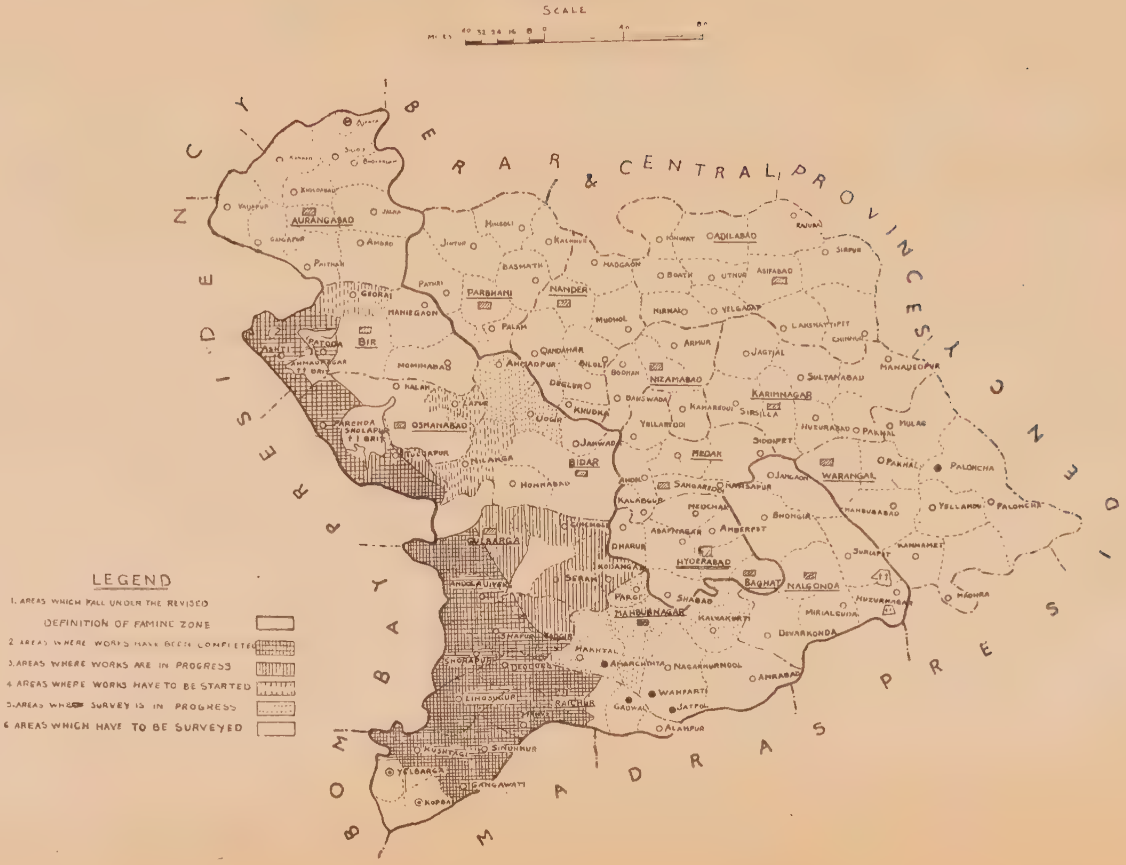
PLATE II

ACTIVITIES

of

THE WELL SINKING DEPARTMENT

HYDERABAD


SCALE
MILES 40 32 24 16 8 0 40 80
BERAR & CENTRAL PROVINCES
MADRAS PRESIDENCY
BOMBAY PRESIDENCY
AURANGABAD
BIR
PARBHANI
NANDER
ADILABAD
NIZAMABAD
KARIMNAGAR
WARANGAL
MEDAK
BIDAR
OSMANABAD
GULBARGA
HYDERABAD
BAGHAT
NALGONDA
MAHBUBNAGAR
RAICHUR
LEGEND
1. AREAS WHICH FALL UNDER THE REVISED DEFINITION OF FAMINE ZONE
2 AREAS WHERE WORKS HAVE BEEN COMPLETED
3. AREAS WHERE WORKS ARE IN PROGRESS
4 AREAS WHERE WORKS HAVE TO BE STARTED
5. AREAS WHERE SURVEY IS IN PROGRESS
6 AREAS WHICH HAVE TO BE SURVEYED

A FEW WELLS IN RAICHUR DISTRICT PLOTTED TO POSITIONS SHOWING THE SUBSOIL STRATA THROUGH WHICH THEY ARE SUNK.

SCALE VERTICAL

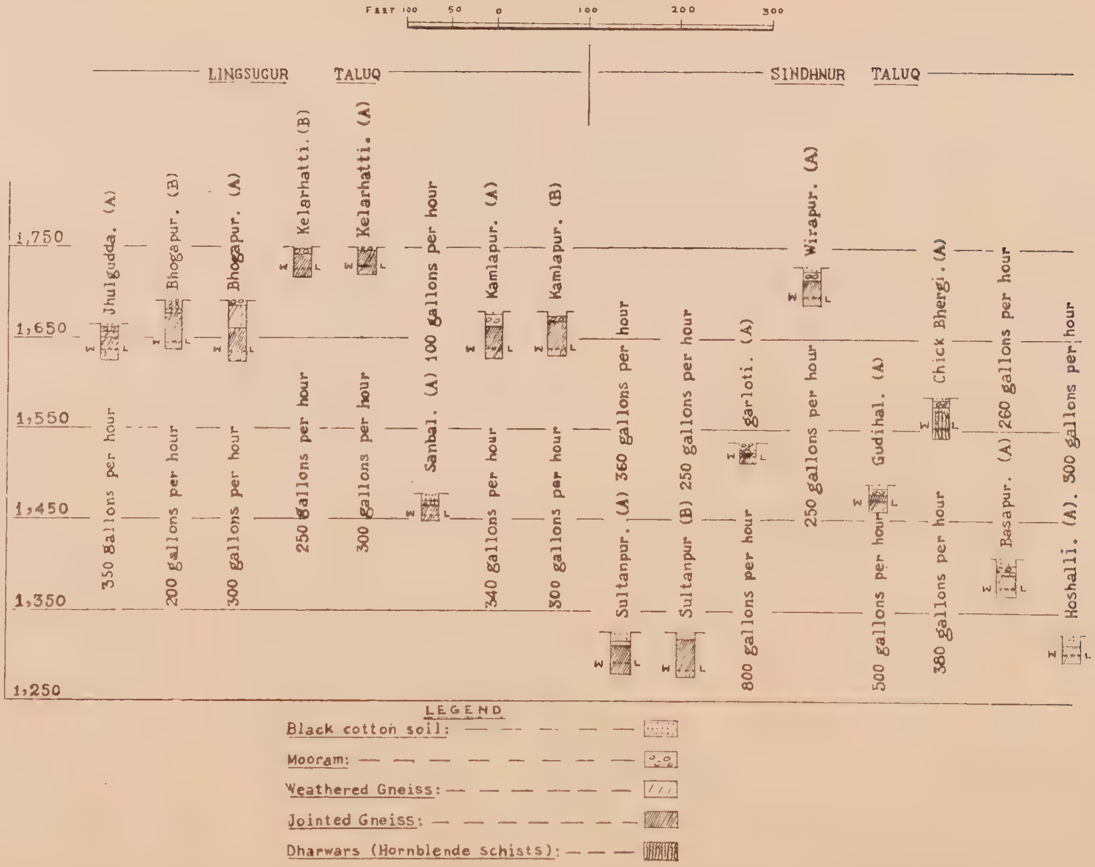
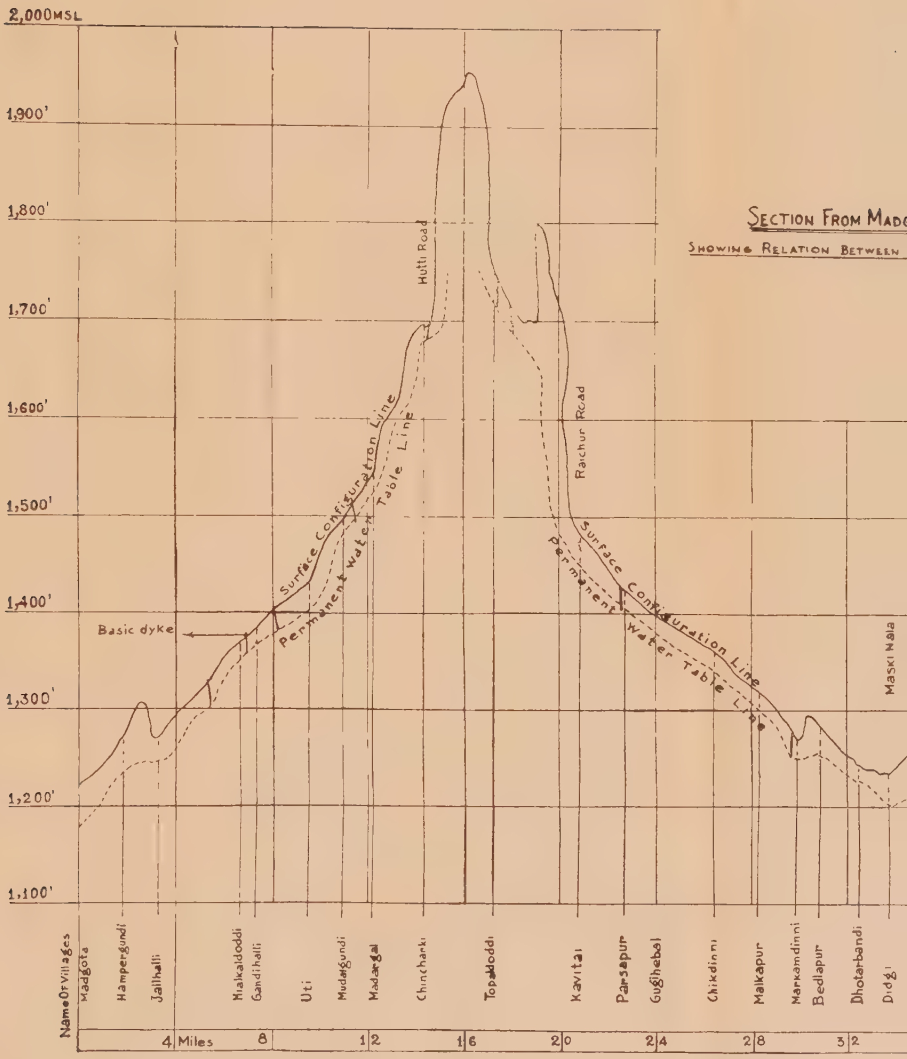

FEET 100 50 0 100 200 300
LINGSUGUR TALUQ
SINDHNUR TALUQ
1,750
1,650
1,550
1,450
1,350
1,250
Jhulgudda. (A) 350 gallons per hour
Bhogapur. (B) 200 gallons per hour
Bhogapur. (A) 300 gallons per hour
Kelarhatti. (B) 250 gallons per hour
Kelarhatti. (A) 300 gallons per hour
Sanbal. (A) 100 gallons per hour
Kamlapur. (A) 340 gallons per hour
Kamlapur. (B) 300 gallons per hour
Sultanpur. (A) 360 gallons per hour
Sultanpur (B) 250 gallons per hour
garloti. (A) 800 gallons per hour
Wirapur. (A) 250 gallons per hour
Gudihal. (A) 500 gallons per hour
Chick Bhergi. (A) 380 gallons per hour
Basapur. (A) 260 gallons per hour
Hoshalli. (A). 300 gallons per hour
M
L
LEGEND
Black cotton soil: ⬚
Mooram: ⬚
Weathered Gneiss: ▨
Jointed Gneiss: ▨
Dharwars (Hornblende schists): ▥

PLATE III

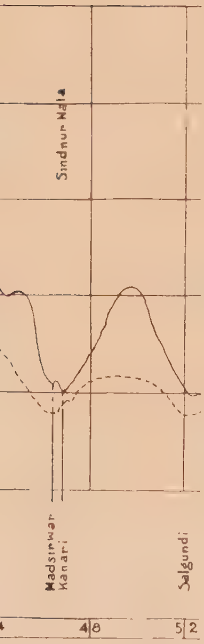

SECTION FROM MADG
SHOWING RELATION BETWEEN
2,000 MSL
1,900'
1,800'
1,700'
1,600'
1,500'
1,400'
1,300'
1,200'
1,100'
Hutti Road
Raichur Road
Surface Configuration Line
Permanent water Table Line
Surface Configuration Line
Permanent Water Table Line
Basic dyke
Maski Nala
Name Of Villages
Madgota
Hampergundi
Jallhalli
Mialkaldoddi
Gandihalli
Uti
Mudalgundi
Madargal
Chincharki
Topaldoddi
Kavitai
Parsapur
Gugihebal
Chikdinni
Malkapur
Markamdinni
Bedlapur
Dhotarbandi
Didgi
4
Miles
8
12
16
20
24
28
32

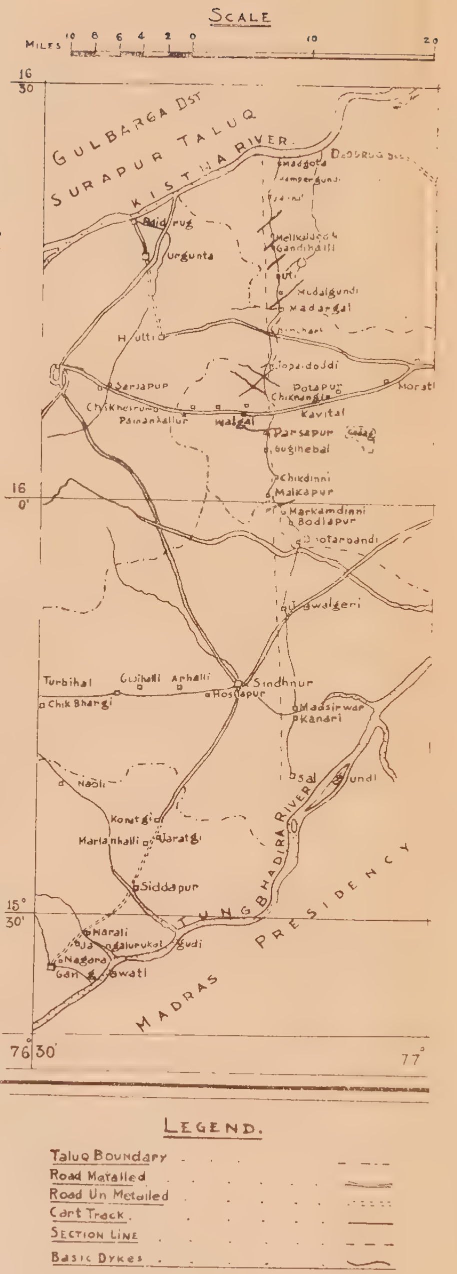
Of Raichur District

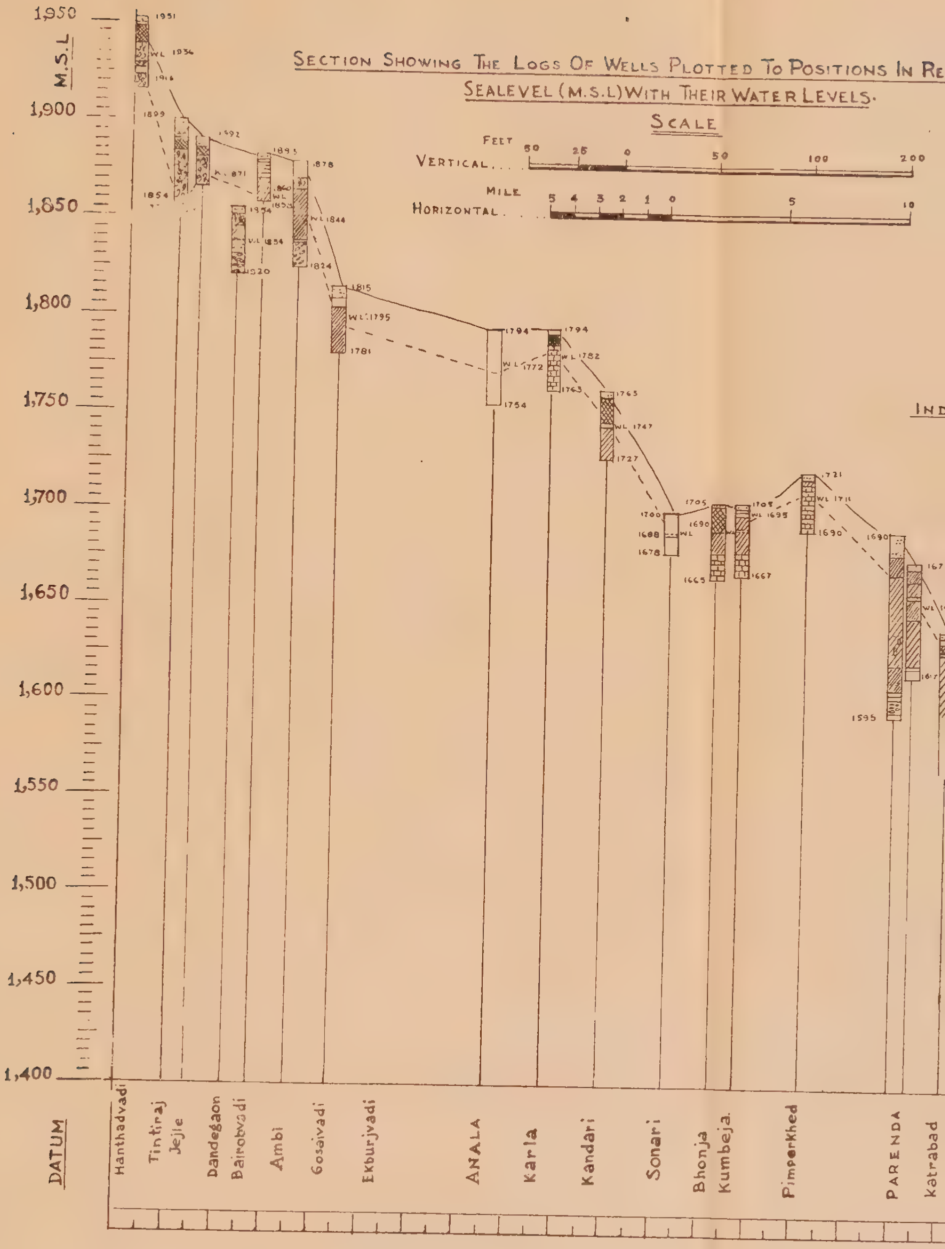
ection Line

o Salgundi

o Surface Configuration



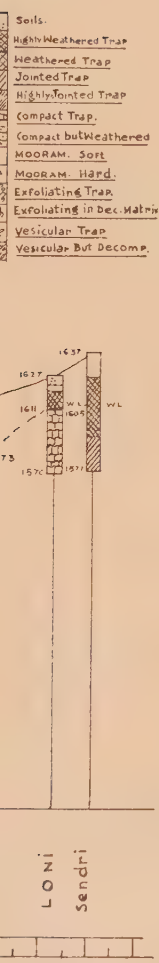




SECTION SHOWING THE LOGS OF WELLS PLOTTED TO POSITIONS IN RE
SEALEVEL (M.S.L) WITH THEIR WATER LEVELS.

EAN

PLATE VI


Soils.
Highly Weathered Trap
Weathered Trap
Jointed Trap
Highly Jointed Trap
Compact Trap.
Compact but Weathered
MOORAM. Soft
MOORAM. Hard.
Exfoliating Trap.
Exfoliating in Dec. Matrix
Vesicular Trap
Vesicular But Decomp.
1637
1627
1611
WL
1605
WL
1570
1577
Loni
Sendri

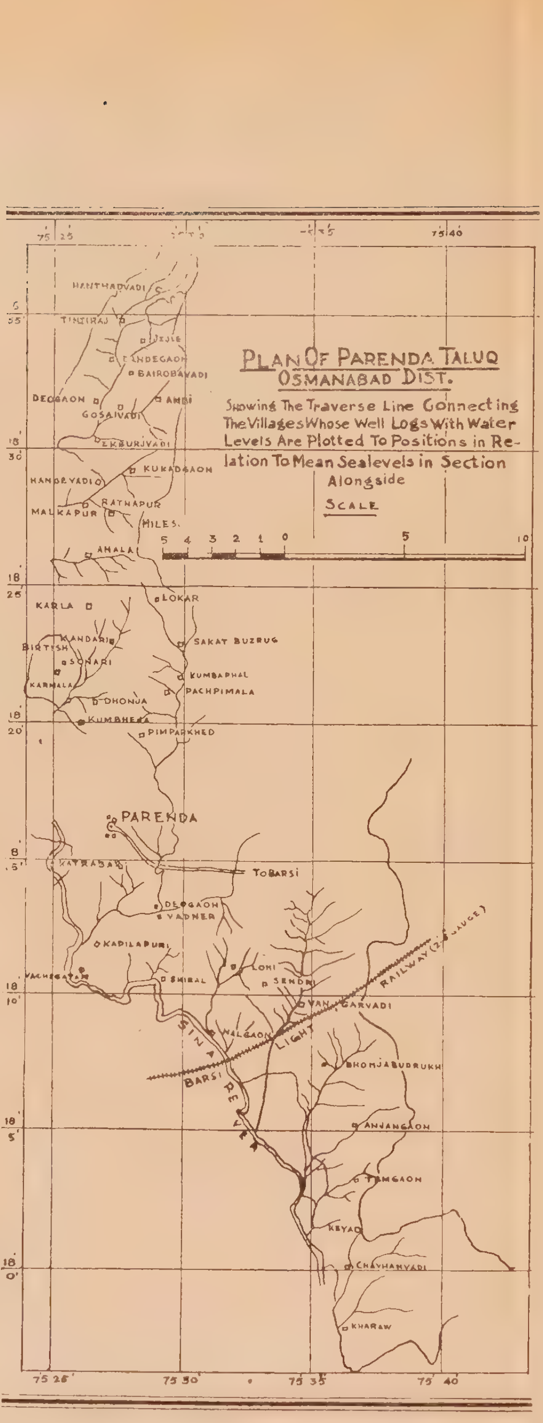
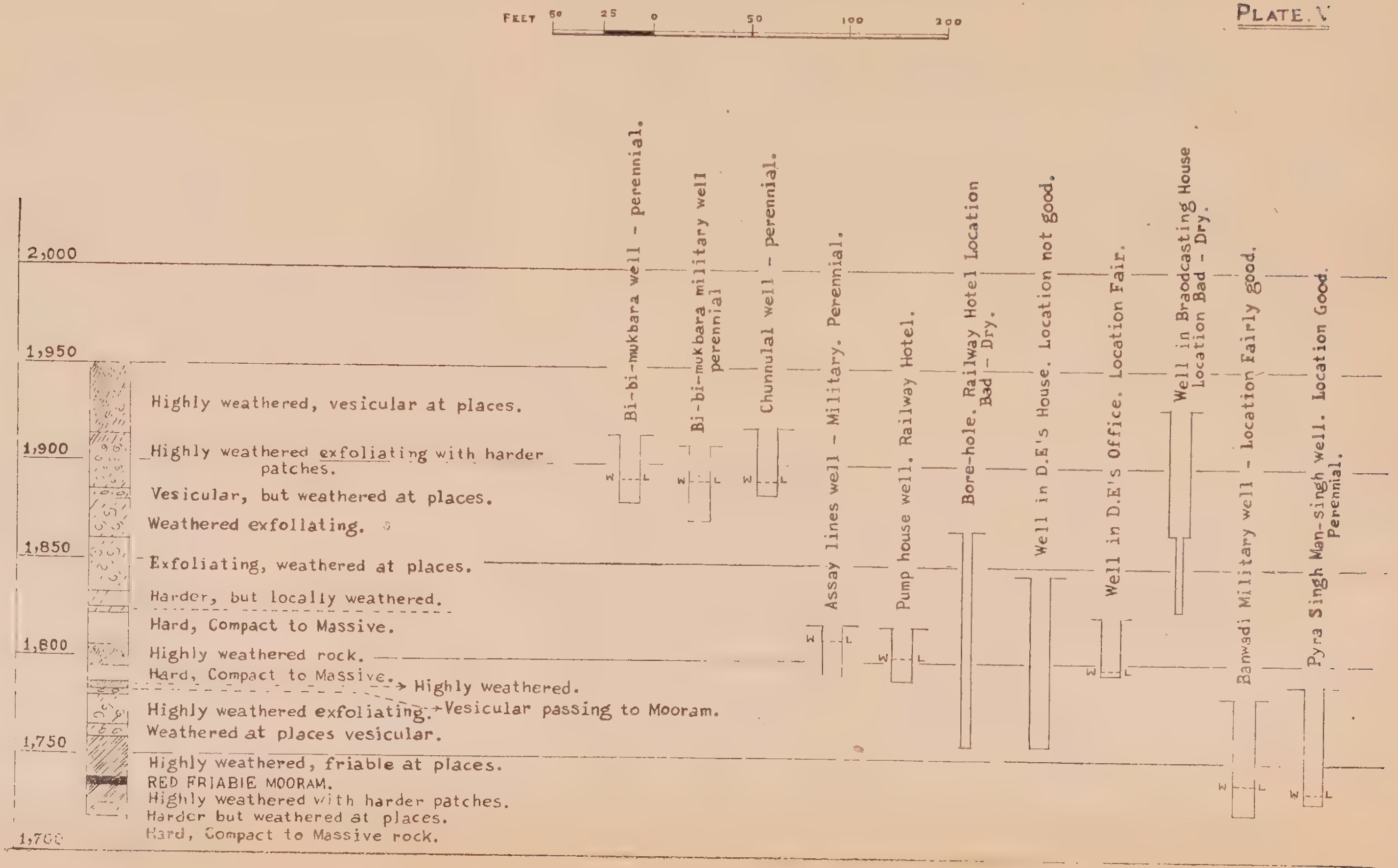

PLAN OF PARENDA TALUQ
OSMANABAD DIST.
Showing The Traverse Line Connecting The Villages Whose Well Logs With Water Levels Are Plotted To Positions in Relation To Mean Sealevels in Section Alongside
SCALE
MILES.
5 4 3 2 1 0 5 10
75 25
75 30
75 35
75 40
18 35'
18 30'
18 25'
18 20'
18 15'
18 10'
18 5'
18 0'
HANTHADVADI
TINTIRAJ
JEJLE
KANDEGAON
BAIROBAVADI
DEOGAON
AMBI
GOSAIVADI
EKBURJVADI
KUKADGAON
HANDEVADI
RATNAPUR
MALKAPUR
AHALA
LOKAR
KARLA
SAKAT BUZRUG
KANDARI
BIRTISH
SONARI
KUMBAPHAL
KARMALA
PACHPIMALA
DHONJA
KUMBHEJA
PIMPARKHED
PARENDA
KATRABAD
TO BARSI
DEOGAON
VADNER
KADILAPURI
VACHEGAVAN
SHIRAL
LONI
SENDRI
VAN
GARVADI
RAILWAY (2' GAUGE)
WALGAON
LIGHT
SINA
BARSI
RIVER
BHOMJA BUDRUKH
ANJANGAON
TAMGAON
KEYAD
CHAVHANVADI
KHARAW

PLATE. V

SEQUENCE OF TRAPS AT AND AROUND AURANGABAD AND THE WELLS THEREIN PLOTTED TO POSITIONS.

SCALE, VERTICAL.


FEET 50 25 0 50 100 200
2,000
1,950
1,900
1,850
1,800
1,750
1,700
Highly weathered, vesicular at places.
Highly weathered exfoliating with harder patches.
Vesicular, but weathered at places.
Weathered exfoliating.
Exfoliating, weathered at places.
Harder, but locally weathered.
Hard, Compact to Massive.
Highly weathered rock.
Hard, Compact to Massive. → Highly weathered.
Highly weathered exfoliating. → Vesicular passing to Mooram.
Weathered at places vesicular.
Highly weathered, friable at places.
RED FRIABIE MOORAM.
Highly weathered with harder patches.
Harder but weathered at places.
Hard, Compact to Massive rock.
Bi-bi-mukbara well - perennial.
Bi-bi-mukbara military well perennial
Chunnulal well - perennial.
Assay lines well - Military. Perennial.
Pump house well. Railway Hotel.
Bore-hole. Railway Hotel Location Bad - Dry.
Well in D.E's House. Location not good.
Well in D.E's Office. Location Fair.
Well in Braodcasting House Location Bad - Dry.
Banwadi Military well - Location Fairly good.
Pyra Singh Man-singh well. Location Good. Perennial.
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2) GROUND WATER IN NORTHERN CHILE — A SUMMARY

by

George C. TAYLOR, Jr.

Abstract

In much of northern Chile ground water is the only source of water for domestic and stock needs, for municipal and industrial supplies, and for irrigation. For many years ground water for irrigation has been obtained by gravity from drains, infiltration galleries, or springs. Economic necessity for increased food production in northern Chile has stimulated national interest in extending the development of ground water for irrigation by pumping from wells.

Throughout northern Chile arid or semiarid climate prevails. Precipitation is scanty or almost entirely lacking in areas with elevations below 2,000 meters. However, in the Andean highlands regular precipitation occurs — the average annual amount varying chiefly with elevation and latitude.

In the region lying between 26° and 33° south latitude, or the so-called « Norte Chico », the most important ground-water bodies occur in the alluvial deposits of the major stream valleys. These bodies are recharged chiefly by infiltration from (1) the floods of winter rains from May to September or (2) the runoff of melt water from Andean snowfields during the spring, that is, from September through November. The ground water discharges in springs that rise along the stream channels near the coast or in submarine springs.

In the region lying between 17° 30′ and 26° south latitude, or the so-called « Norte Grande », extensive ground-water bodies occur in alluvial deposits that fill basins of interior drainage such as the Pampa del Tamarugal and the Atamaca basin Locally in or near these basins productive aquifers occur in Miocene sandstones and conglomerated intercalated with volcanic rocks. To the east of these basins, in the Andean highlands, short but intense thunderstorms are common during the summer months, that is, from December to March. The recharge of the ground-water bodies takes place by infiltration from the accumulation of such flood waters. The ground water discharges by (1) evaporation from the capillary fringe or the water table where it is near the surface or by (2) transpiration of phraetophytic plants in areas where the water table is at shallow depth. Such discharge occurs in and near the salt flats that occupy the lower parts of the principal basins of interior drainage.

Introduction

During 1946-47 the writer was assigned by the United States Geological Survey to work with the Irrigation Section of the Chilean Development Corporation (Corporacion de Fomento de la Produccion de Chile) in a ground-water survey of the northern part of the republic. The present paper describes briefly some general aspects of the geology and ground-water hydrology of the region between Santigoa and the Peruvian frontier.

Geography

Northern Chile occupies a strip some 100 to 350 kilometers wide and 1,600 kilometers long. It lies roughly parallel to the meridian 70° west longitude, and between 17° 30′ and 33° south latitude. To the east it is limited by the great mountain chain of the Andes Range, and to the west by the Pacific Ocean. That part of the country lying between 62° and 33° south latitude is known in Chile as the « Norte Chico » (« Little North »), and the part that extends from 26° to the Peruvian frontier is called « Norte Grande », (« Big North »).

The « Norte Chico » is largely a rugged, mountainous highland that rises steeply from the Pacific coast to elevations of some 4,000 to 4,500 meters along the crest of the Andes 100 to 200 kilometers inland. The principal fluvial systems of the « Norte Chico » are those of the Ríos Copiapó, Huasco, Los Choros, Elqui, Limari,

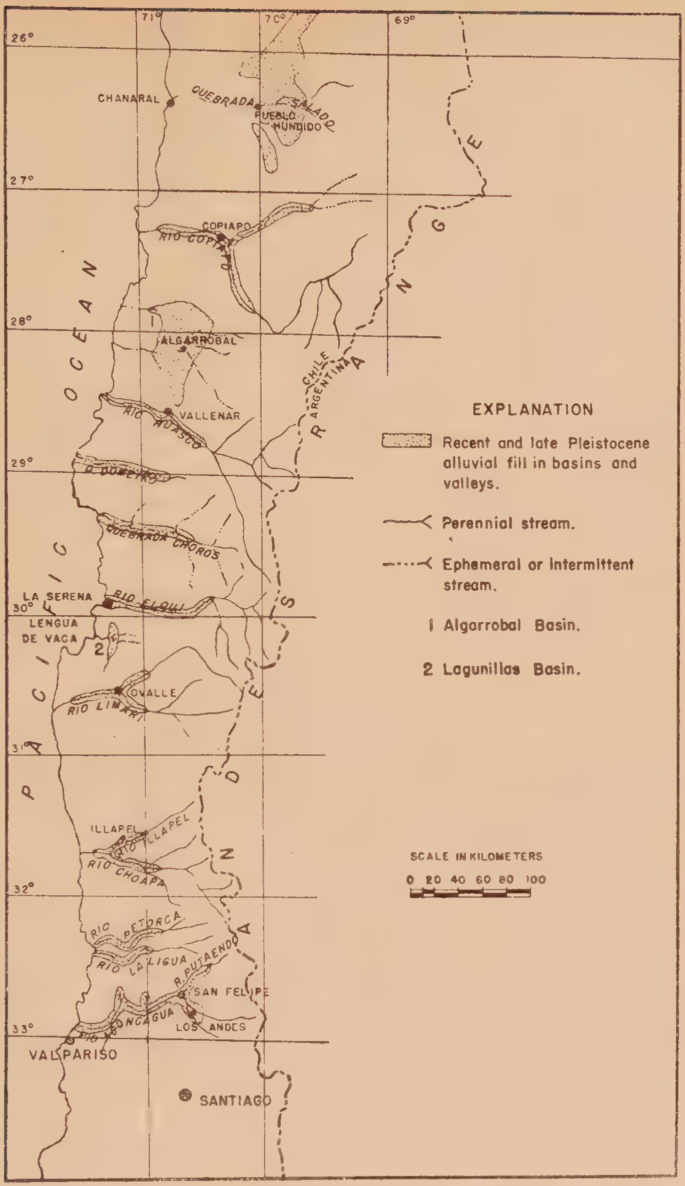

Fig. 1.

Choapa, Petorca, La Ligua, and Aconcagua (Fig. 1). These streams and their tributaries rise on the western slopes of the Andes and flow to the Pacific through deep gorges and narrow, winding valleys cut through the highlands of the region. Near the sea the streams are bordered by fluvial terraces that may extend inland for as much as 30 to 40 kilometers.

The « Norte Grande » (fig. 2) may be divided into three regions of distinct topographic aspect. To the west are the Coast Ranges that rise abruptly from the sea in a great escarpment 500 to 800 meters high. The Coast Ranges form an irregular highland belt some 40 to 70 kilometers wide. The average elevations of this belt

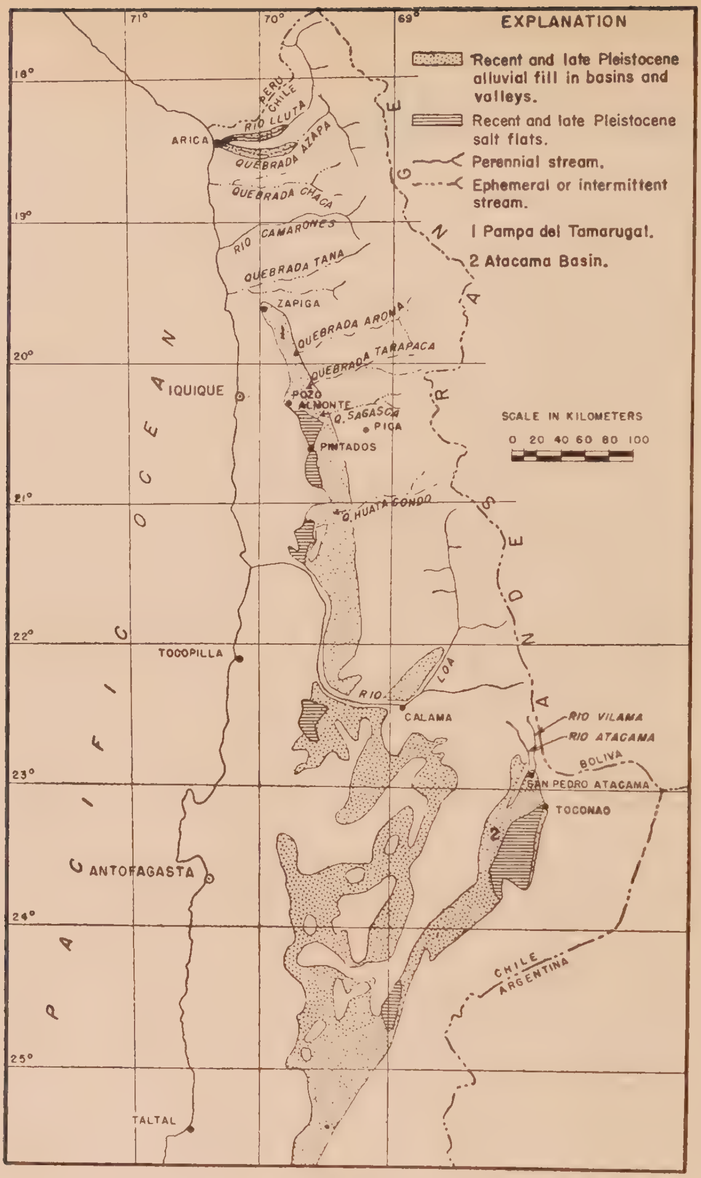

Fig. 2.

range between 1,000 and 1,500 meters, and the maximum elevations do not greatly exceed 2,000 meters. Lying between the Coast Ranges and the Ranges is a north-south or longitudinal depression 25 to 50 kilometers wide. This depression generally lies between 900 and 1,300 meters above sea level. In the southern part of the « Norte Grande » its surface is broken into smaller basins or pampas by isolated groups of hills or spurs of the Coast Ranges and the Andes. Approximately in the region of 26° south latitude the longitudinal depression disappears, and spurs of the Coast Ranges and of the Andes merge in the highlands of the « Norte Chico ». Throughout much of the « Norte Grande » the western flank of the Andes Range is a monoclinal slope. This rises eastward from the longitudinal depression and culminates in a broad, undulating plateau some 4,000 meters above sea level. Volcanoes built up on the surface of this plateau reach elevations of more than 5,000 meters in some places. Because of the scanty precipitation in the « Norte Grande » most of the streams are ephemeral, and many lose themselves in bolsons or basins of interior drainage that are characteristic of the region. The only permanent streams of the « Norte Grande » are the Rios Lluta, Camarones, and Loa. The waters of all of these are brackish and therefore of limited value for irrigation or other beneficial use.

Precipitation

In practically all parts of northern Chile, arid or semiarid climatic conditions prevail. That part of the « Norte Chico » to the south of 29° south latitude lies in the belt of prevailing westerly or northwesterly winds and receives periodic rains during the winter months, from May to September. This belt shifts with the seasons and makes itself felt as far north as 27° south. At Copiapó the mean annual rainfall is about 15 millimeters; at La Serena 160 millimeters; and at Illapel 250 millimeters. Precipitation in this region also increases with altitude. In the same latitude the higher slopes of the Andes may receive 3 to 5 times more precipitation than the areas near sea level. Snow falls are common at altitudes above 2,000 meters during the cold months.

In all of the « Norte Grande » below elevations of about 2,000 meters precipitation occurs at such infrequent intervals that it may be said that practically it never rains. However, periodic precipitation occurs in the Andes Range at elevations above 2,000 meters. This increases with altitude so that at 4,000 meters the annual precipitation is of the order of 250 to 300 millimeters. Practically all precipitation falls in short but intense convectional storms of rain, hail, or snow during the summer months, from December through March. These storms are of local character and sporadic distribution; hence the annual rainfall at any one station may vary 300 per cent or more from year to year.

General Geology

Viewed in broad outline the geology of northern Chile is relatively simple. The oldest rocks are probable Paleozoic or pre-Cambrian age. They occur in limited areas along the coast of the « Norte Chico » chiefly near the mouth of the Rîos Choapa and Petorca, and in the peninsula of Lengua de Vaca. They include slate, quartzite, tillite, and granite that have been considerably metamorphosed.

Triassic rocks also appear along the coast between the Rîos Choapa and Petorca, and between the Rios Huasco and Copiapó. These rocks are chiefly sediments of continental origin. They also include marine sandstone, shale, and limestone with intercalated beds of acid lavas.

The most widespread group of rocks in northern Chile is of Jurassic age. Most of these rocks belong to the so-called « Formación Perfirîtica », a heterogeneous assemblage of andesite porphyry, conglomerates, breccias, and tuffs. In the lower part of the formation marine limestones and shales of Middle and Lower Jurassic age are intercalated with the volcanic rocks. The Jurassic rocks pass without pronounced stratigraphic break into marine limestones and shales of Lower Cretaceous age.

In the Late Cretaceous the rocks of northern Chile were uplifted and deformed, chiefly in the Andean region. Accompanying the deformation was the intrusion of extensive masses of igneous rock belonging to the so-called Andean diorite. The magma from which these intrusive masses originated apparently was largely dioritic in composition, but extensive local differentations of granite, granodiorite, and even gabbro occur. These rocks crop out extensively along the coastal zone of the « Norte Chico » to the south of 26° south latitude but appear only in small intrusive masses in the « Norte Grande ».

In early Tertiary time the Cretaceous and older rocks were dislocated by faults — chiefly along north-south lines. As a result of these movements numerous closed depressions separated by fault blocks were developed. Also, the great longitudinal depression of the « Norte Grande » must have originated as a graben between the Coast Fanges and the Andes at this time. Because of the higher rainfall and greater competence of streams in the « Norte Chico », most of the larger rivers of this region were able to reestablish external drainage to the sea following the fault movements.

However, the ephemeral streams of the « Norte Grande » were not capable of carrying their sediment loads to the sea but instead deposited them in basins of interior drainage between the fault ridges.

In the middle Tertiary, probably in Miocene time, widespread volcanic eruptions occurred in the Andean region of the « Norte Grande ». These eruptions took place largely from centers located along the eastern frontier of Chile. In the Andean region enormous quantities of rhyolitic and dacitic lava, tuff, and breccia were erupted. At the same time Andean streams filled the longitudinal depression with detrital deposits of soft sandstone, siltstone, and shale, and some gravel. This heterogeneous assemblage of volcanic and sedimentary rocks, known collectively as the « Formación Liparitica », underlies large areas of the « Norte Grande ».

Slight movements of the continental margin occurred during Pliocene time, and in small areas of the « Norte Chico » marine waters encroached inland short distances from the present coast line. In these areas fossiliferous limestones, sand, and clay were deposited. At the same time the principal streams of this regions aggraded the coastal sections of their valleys with alluvial deposits of gravel, sand, and silt.

Uplift on a continental scale affected most of northern Chile toward the end of Pliocene time. In the « Norte Grande » the Andes Range was uplifted to approximately its present height in a great tectonic flexure. At the same time the continental margin sank below sea level along the fault line marked by the present coastal escarpment of northern Chile.

During Pleistocene time Andean streams of the « Norte Grande » continued to deposit their sediments in the longitudinal depression. However as a consequence of the late Pliocene uplift, the Rîo Loa and the streams north of 19° 30′ that had previously established outlets to the sea cut deep valleys through the Coast Ranges and across the longitudinal depression. The quebradas (canyons) of Rîos Lluta, Azapa, Chaca, Camarones, and Tana were formed in this manner. The principal drainage systems of the « Norte Chico » had probably been established in middle Tertiary time. However, as a consequence of the late Pliocene uplift these streams deepened their valleys by cutting down into their Pliocene alluvial deposits, and also into the older consolidated rocks.

In the Andean region considerable volcanic activity took place at numerous eruptive centers north of latitude 27° during Pleistocene time. Eruptions of intermediate and basic lavas at these centers resulted in the building up of the modern volcaneos of the region.

During late Pleistocene and Recent time most of the larger streams of northern Chile aggraded their lower courses with unconsolidated deposits of gravel, sand, and silt. At the same time aggradation continued in basins of interior drainage of the « Norte Grande ».

Water-bearing character of the rocks

The Cretaceous and older rocks of northern Chile do not generally contain important aquifers or water-bearing horizons. They form the dissected highland areas and therefore are usually rather thoroughly drained. Moreover, these rocks are indurated and generally lack primary openings in which ground water might accumulate or be transmitted in important quantities. Some ground water circulates through secondary openings in these rocks, chiefly along joints and other fractures or in surficial mantles of disintegrated rock. Such water may sustain small springs or may supply domestic wells. However, in general the quantities of water available from these rocks are not sufficient for large-scale use.

The Miocene « Formación Liparîtica » contains important aquifers in certain favorable areas of the « Norte Grande ». The most productive aquifers are permeable sandstones and conglomerates that are intercalated with impervious volcanic and sedimentary strata. In the Pica zone (latitude 20° 30′ south) three to five such aquifers occur between depths of 25 and 250 meters. Deep water-bearing sandstones and conglomerates also occur in the Pampa del Tamarugal (between 19° 31′ and 21° south, fig. 2). These aquifers in the « Formación Liparîtica » have been encountered between depths of 136 and 315 meters by wells drilled near the Pintados salt flat. Water in these aquifers is confined, but the artesian pressure generally is not sufficient to cause wells to overflow at the surface. On the western side of the Atacama Basin (between latitudes 23° and 24° south) are several artesian springs that rise from aquifers in the « Formación Liparîtica ».

Along the lower courses of the major streams of the « Norte Grande » are Pliocene fluvial deposits of clay, sand, and gravel that grade into marine limestone, clay, and sand along the coast line. These deposits have been cut by stream and marine erosion into terraces that rise 10 to 60 meters above present drainage levels. The sand and gravel members of the deposite are moderately permeable and may yield small quantities of water to springs or domestic wells. However, because of their dissection the deposits are generally rather thoroughly drained.

The most productive aquifers in northern Chile occur in late Pleistocene and Recent alluvial deposits that fill the lower parts of the basins of interior drainage in the « Norte Grande » (fig. 2) and the principal stream valleys in the « Norte Chico » (fig. 1). These deposits consist of intercalated beds of clay, silt, sand, and gravel laid down by ephemeral, intermittent, or perennial streams. Where present in the zone of saturation, the deposits may yield moderate to large supplies of water to wells and springs, depending largely on local permeability. The generall occurrence of ground water in the alluvial deposits is discussed below.

Occurrence of ground water in the alluvial deposits

Norte Chico

In the « Norte Chico » late Pleistocene and Recent alluvial deposits occur along the middle and lower courses of the principal stream valleys. These deposits lie in narrow but continuous bands generally ranging in width from a few tens of meters to perhaps 3 kilometers. However, along the Rio Aconcagua and its principal tributaries, the deposits underlie alluvial valleys from 5 to as much as 16 kilometers wide. The deposits are generally less than 50 meters thick, but in places may be more than 100 meters thick.

The ground water in the late Pleistocene and Recent alluvial deposits originates from precipitation within the drainage basin of the principal streams of the « Norte Chico ». The alluvial deposits are composed largely of unconsolidated gravel and sand that is moderately to highly permeable. In each of the larger alluvial valleys a part of the water that falls as precipitation in the drainage basin accumulates in these deposits and forms a zone of saturation. The recharge of water to this zone takes place largely by infiltration from streams. Such recharge occurs principally during (1) floods resulting from winter rains, from June to September, or (2) the runoff of melt water from Andean snowfields in the spring months, from September to November.

The upper limit of the zone of saturation in each of the valleys of the « Norte Chico » is marked by a water table. This fluctuates seasonally in response to the replenishment or depletion of water in the zone of saturation. The highest stages of the water table generally occur in late spring, that is, in October or November. The lowest stages occur in the late fall, that is, April or May, near the end of the dry season. For example, in the valley of the Río Putaendo the water table may fluctuate 3 to 4 meters as a result of seasonal changes in ground-water storage. Comparable fluctuations of the water table may occur in the other valleys of « Norte Chico ».

The water table in each of the principal valleys slopes downstream with a gradient approximately parallel to that of the surface stream channel. The water accumulated in the zone of saturation moves down the slope of the water table. In valleys of the « Norte Chico » occupied by perennial streams such as the Rios Huasco, Elqui, Limari, Choapa, and Illapel, the elevation of the water table generally is about the same as that of the stream. However, in certain stretches of the valleys of the Rios Copiapó and Aconcagua, the streams flow on impervious bottoms and thus are insulated, and the water table may be several meters or more below the surface. The Rios Petorca and La Ligua are both intermittent in their middle and lower courses. Characteristically the perennial spring-fed stretches are separated by dry stretches that carry surface water only from flood runoff. The dry-season flow in the perennial stretches may be of the order of 50 to 500 litres per second. In the intervening dry stretches of these streams the water table is generally less than 5 meters below the surface. Except for short spring-fed stretches near the sea the Quebradas of Domeyko and Los Choros are ephemeral. In the dry stretches of the alluvial valleys of these quebradas, the water table may range from a few meters to as much as 50 meters below the surface.

Because of the limited witdh of the late Pleistocene and Recent alluvial deposits in valleys of the « Norte Chico », the ground water moves essentially in underflow conduits. Very rough estimates based on empirical observations suggest that the underflow in the lower valley of the Río Putaendo may be about 1 ½ to 2 cubic meters per second. The valleys of the Ríos Petorca and La Ligua are each estimated to have underflows of the order of 500 to 800 liters per second. The underflow in the lower part of the Quebrada of Los Choros may be no more than 300 liters per second, and that of the Quebrada of Los Domeyko, no more than 75 liters per second. The ground water that moves through the underflow conduits discharges in spring. along the streams near the coast or emerges below sea level in submarine springs.

Norte Grande

In the « Norte Grande » late Pleistocene and Recent alluvial deposits occur most extensively in the Pampa del Tamarugal, in the Atacama basin, and in smaller basins of interior drainage near the southern end of the longitudinal depresssion (fig. 2). These deposits underlie most of the Pampa del Tamarugal over an area about 210 kilometers long by 30 to 40 kilometers wide. They also underlie the Atacama basin in an area about 110 kilometers long by 65 kilometers wide (fig. 2). In the Pampa del Tamarugal these deposits may attain a thickness of 100 meters, but in the Atacama basin the thickness is not known even in general terms.

The late Pleistocene and Recent deposits of the Pampa del Tamarugal are made up largely of great interlocking alluvial fans of Andean quebradas. Some of these fans extend completely across the pampa to the base of the Coast Ranges. To the east, near the heads of the fans, the alluvial deposits are coarse gravels with sand. Farther to the west the deposits grade progressively into sand with fine gravel, and finally, near the Coast Range border, the deposits consist chiefly of intercalated lenses of silt, clay, and fine sand.

The ground water that occurs in the late Pleistocene and Recent alluvial deposits of the Pampa del Tamarugal is originally derived from precipitation on the Andes Range at elevations above 2,000 meters. Most of the annual precipitation occurs in convectional storms of local character during the months December through March. The runoff from such storms is very rapid and passes down the larger Andean quebradas in flash floods of large volume but short duration. Upon reaching the pampa the flood waters leave the stream channels and spread out in great sheets of water that may inundate many square kilometers.

The ground-water body beneath the Pampa del Tamarugal is sustained and replenished almost entirely by the infiltration from such flood waters. The larger Andean quebradas also contribute considerable quantities of ground water to the pampa by the underflow through their alluvial deposits. For example, the Quebrada of Tarapacâ is estimated to contribute a perennial underflow of about 250 liters per second.

In general, the upper limit of the zone of saturation in the pampa is marked by a water table. This slopes west-southwest from the base of the Andes to the Coast Range border. It also slopes south from Negreiros to the salt flats of Pintados and Bella Bista (fig. 2). In these areas natural ground-water discharge takes place by (1) direct evaporation from the water table or from the capillary fringe and (2) by the transpiration of phreatophytic plants such as tamarugos, algarrobos, and salt grass.

Near the salt flats of Pintados and Bella Vosta and near Zapiga, the water table is generally within 25 meters of the surface. However, in intervening areas of the pampa and within a few kilometers of the Coast Range border, the water table lies between depths of 25 and 75 meters. Along the eastern side of the pampa the water table generally lies deeper than 75 meters.

Ground-water conditions in the Atacama basin (fig. 2) are generally similar to the Pampa del Tamarugal. The principal source of ground water in the late Pleistocene and Recent deposits of the Atacama basin is from prepicitation on the high plateau to the east and from the infiltration of flood waters of the Rios Vilama and Atacama. The natural discharge of ground water in the basin occurs chiefly through direct evaporation from the water table or from the capillary fringe in the great Atacama salt flat. (fig. 2)

Utilization of ground water

Throughout much of northern Chile the principal source of water supply for domestic, stock, municipal, industrial, and irrigation needs is ground water obtained from springs, wells, or infiltration galleries. The most productive water-bearing beds occur in late Pleistocene and Recent alluvial deposits of sand and gravel. Locally, however, Miocene rocks contain productive water-bearing beds of sandstone or conglomerate.

The most important areas of ground-water development in the « Norte Grande » are the Azapa valley, the Pampa del Tamarugal, the Pica zone, and to lesser extent the Atacama basin (fig. 2). In the Azapa valley water-bearing beds of late Pleistocene and Recent sand and gravel occur at depths of 2 to 35 meters. These crop out at the surface and support springs or have been developed artificially by infiltration galleries or wells. In recent years most of the springs and infiltration galleries have gone dry, and wells have been used to an increasing degree. At present there are some 40 active irrigation wells in the lower 30 kikometers of the valley. The dug wells range from 15 to 30 meters deep and yield from 6 to 12 liters per second to pumps. Somewhat higher yields have been obtained in drilled wells put down by the Department of Irrigation of the Bureau of Public Works. The most productive well in the valley was drilled to a depth of 36 meters. It yields 32 liters per second with a deep-well turbine pump. Ground water in the Azapa valley is developed principally for domestic purposes, for the irrigation of olive and citrus groves, and for the public supply of Arica.

The late Pleistocene and Recent deposits of the Pampa del Tamarugal contain productive water-bearing strata of sand and gravel. Along the western side of the pampa were, during the flourishing epoch of the nitrate industry, some 160 active wells that tapped these strata. The wells were dug to depths ranging from 2 to 85 meters and were pumped individually at rates up to 500 cubic meters per day. The water was used principally for processing nitrate ore and to a minor extent for the irrigation of garden plots belonging to the nitrate companies. The quality of the water from the wells ranges from salty or brackish to fresh. Most of these wells are now idle or abandoned.

Irrigation wells drilled in the vicinity of the Pintados salt flat have encountered water-bearing sandstones of probable Miocene age between depths of 136 and 315 meters. These wells yield from 15 to 80 liters per second with deep-well turbine pumps. The water is generally of good quality.

In the Pica zone important water-bearing horizons occur at depths of 25 to 250 meters in Miocene sandstones and conglomerates intercalated with volcanic rocks. The water from these horizons rises to the surface in large artesian springs or is tapped by wells and infiltration galleries under both artesian and water-table conditions. It has been developed for the public supply of Iquique and for the irrigation of some 100 hectares of citrus orchards and other crops in the vicinity of Pica.

Water in the late Pleistocene and Recent deposits of the Atacama basin has been developed only to a limited extent for domestic and stock use, near San Pedro de Atacama. In this vicinity are several dug wells from 5 to 25 meters deep that are equipped with windmills or bucket and windlass. Near Toconao is a small oasis irrigated by artesian springs that rise from aquifers of the Miocene « Formación Liparítica ».

In the valleys of the « Norte Chico » water in late Pleistocene and Recent alluvial deposits is used extensively for domestic and stock requirements. It has been developed by dug wells ranging from a few meters to 50 meters deep. In the valley of the Río Aconcagua a large infiltration gallery being constructed in the alluvial deposits is expected to produce some 500 liters per second for the public water supply of Valparaiso. Irrigation by gravity diversion of springs that rise from the alluvial deposits is practiced in all the major valleys of the « Norte Chico ». However, pumping for irrigation from wells in these deposits has not been generally feasible because of the adverse economic factors prevailing in the region.

3) REPORT ON THE GROUND WATER CONDITIONS OF THE GREAT HUNGARIAN PLAIN

by

Doz. Dr. JOHN BOGÁRDI
director of the Hungarian Hydrographic Institute

One of the peculiar geographic unites of Europe is to be found the lowland on the bottom of the Carpathian basin extending for about 100.000 square km. originating in the sinking of part of the so-called « Tisia massivum ». The deeply sunk parts were entirely covered by the layers of the third age tide, which were later concealed by continental deposits and leveled to a plain-land. This report is limited to the part of the « Alfôld » (: Great Hungarian Plain). of present-day Hungary, a territory of around 45.000 square km used mainly for agriculture. (1)

The average elevation of Hungarian « Alfôld » above the sea-level (:Adria:) varies between 90 and 130 m. The territory is slightly sloping to the south. In places, as for instance between the Danube and the Tisza, sandhills and mounds of moderate height are present. Elevation differences reach, however, only exceptionally 20 to 40 m.

The Hungarian « Alfòld » belongs in its entirity to the drainage-area of the Danube. Besides the Danube, forming the western boundary, its most important river in the Tisza. The foremost tributaries of the Tisza are the Szamos, Bodrog, Sajó, Körös and Maros. The Danube is in general characterized by low stages in the fall and winter, while floods and lasting high stages occur mostly in the spring and summer. The stage of the Tisza is as a rule lowest in summer. The great floods, however, occur also on the Tisza usually in the spring. The drainage area of the rivers are detached from each other by low, natural elevations, characterized by a comparative permanency. On longer streches upwards from the mouth, however, the boundaries of the drainage areas of the recipient river and tributaries are more or less blurred in accordance with the characteristics of the low-land.

With the exception of the Danube the streams of the Alfôld, including also the Tisza, have extraordinarily variable discharges in accordance with extreme climates. After inclement winters with much precipitation, the discharge of the rivers in the summer is in general small. The climate of the Alfôld is of a continental character. Severe colds in the winter and very hot days in summer are frequent. The annual mean temperature is +10.1 to +11.4 C°. The mean temperatures of the diverse seasons are in order as follows : winter —0.9 to — 0.8 C°, spring +10.7 to +11,7 C°, summer +20.1 to 21.3 C° and fall + 10.5 to +11.9 C°. The annual sunshine-period in the average is 1.952 to 2.072 hours.

The average precipitation for the Alfôld as a whole is about 580 mm. Taking into account the average of many years, the most precipitation appears to have been had in the north-eastern part of the Alfôld where the annual mean was around 700 mm. On the major part of the Alfôld, the annual mean of the precipitation varies between 500-550 mm. Dryest is the Alfôld in the heart of it in the region of city Szolnok where the annual average precipitation is less than 500 mm.

This report concerns itself mainly with questions of the ground water level (main water table), which is important from the agricultural and forestry point of view. This circumstance indicates in itself as to what is meant by ground water in our report. From this point of view *Zunker* expresses best the conception of ground water. He claims that the ground water is a mass of water filling the coherent pores of the soil which is under no pressure following only the law of hydrostatical pressure. Inasmuch as only such water masses are of consequence for the agriculture and forestry, which accumulate in the impervious layer next to the surface our observations and data are also limited to this part only. Being outside of the sphere of the subject we shall refrain from the further discussion of question of the « ground water conception », so much debated in the literature.

The ground water masses are subject to the law of gravitation the same as the surface waters. If their close position is insured by the impervious layers, they must be considered still-waters. But the water table is not horizontal in such cases either, that is the water table does not take its position along a level surface. Similar phenomena occur on homogen soil of minor extent on level surfaces only.

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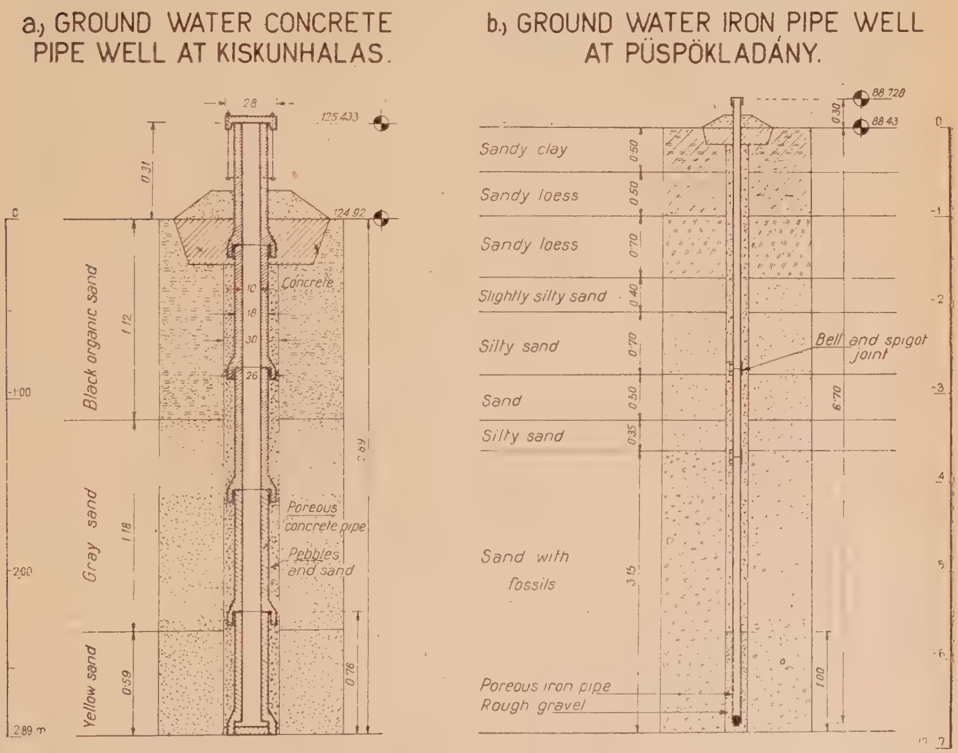
(1) All data contained in the report concern territories located within the boundary of the present-day Hungary.

In general the water table possesses a certain slope and therefore the ground water — similar to the surface water — streams in the direction of the slope. Therefore the ground water stages of certain places are influenced by the infiltration of the precipitation, together with other factors and also side currents. The great soil friction as well as its extreme variation further many other factors delay, however, the fast pressure-difference equalization through streaming in consequence of which the water table very seldom becomes horizontal. *In the general formation of the water table local conditions* play a very considerable role, and so — according to Horusitzky — it has a special topography. Therefore in studying the ground water table, the water surface elevation above the basic level (in our case the level above the Adratic sea) on one hand its elevation or depths relative to the local ground levels on the other hand must be given.

In Hungary the ground water observations were begun very late, about a hundred years later than the hydrographic observations. The first of these observations of experimental character were initiated by the Research Institute of Forestry in 1912. The material of the observations extending to a few wells were, however mostly destroyed without being in use during World War I and the following turbulent changes. Following World War I the forestry service has again started the systematic study of ground water conditions of the Alfôld in accordance with the ideas of prof. Cholnoky in connection with the afforestation of the Alfôld. Observations were now made systematically and continuously. Up to 1928 already the stages of sixtyeight wells were regularly measured. The observation of the ground water conditions on the Alfôld date back in fact to this time.

Ground water studies of the Alfôld took another initiative in 1929. Prof. Rohringer started at this time a study of ground water conditions prevailing in the territory between the Danube and the Tisza. From 1929 to 1934, 149 pipe wells were installed exclusively for the observation of water tables. The wells are lined


a.) GROUND WATER CONCRETE PIPE WELL AT KISKUNHALAS.

b.) GROUND WATER IRON PIPE WELL AT PÜSPÖKLADÁNY.

FIG. 1.

with concrete pipes of ten cm inner diameter. Two or three of the lower pipes are made of porous sandless concrete. Figure 1–a shows the pipe well built in the territory between the Danube and Tisza at Kiskunhalas - Harkapuszta.

The Hydrographic Institute in 1933 also started the development of the well net-work extending over the whole Alföld. In building of the new wells the already existing network was taken into consideration. So far the Hydrographic Institute has set up 220 pipe wells. In connection with some more extensive hydraulic constructions many ground water wells were built by the Hydrographic Institute. Such special purposes for instance, are served by the wells placed along the alignement of the projected Danube-Tisza canal. The ground water wells are built by the Hydrographic Institute with the use of ironpipes of 8 cm inner diameter. On figure 1–b the pipe well built by the Institute at Püspökladány is shown.

It must be mentioned finally that the National Irrigation Office has built since 1938, in connection with the trans-Tisza irrigations, 100 iron wells for the observation of the ground water levels.

The « Rohringer » concrete wells and the iron wells of the National Irrigation Office were gradually taken over by the Hydrographic Institute since 1941 and thus at present the ground water wells are in the majority observed by the Hydrographic Institute. Naturally in the course of time many wells were silted up, but even more numerous are the wells which were damaged and became useless during the recent war. The well network under the management of the Institute consists of 412 pipe wells part of which can be put in operation again after slight or more serious repairs.

The stages in the wells are measured ten times a month or in the average every third day, always downwards from the precisely leveled brim of the well. Thus every stage shows therefore the depth of the water table below the brim of the well.

The orographical development of the Carpathian basin expresses already the hydrographic unity of the region. The hydrographic unity concerns of course also the ground water conditions which is verified by the similar character of *variation of the ground water level all over the country.* Characteristic of the ground water level variation of the Alföld is that at the end of the spring (March, April, May or perhaps June) are the maximums, and in the fall (September, October and November) are the minimums, showing the continental character of the ground water variations of the Alfold.

On account of the climatic conditions of the Alföld, the relatively long and inclement winter prevents the supply of the ground water. The precipitation in form of snow cannot infiltrate in the frozen ground. The supplementing of the ground water can start at the beginning of the melting only, resulting through frequent fast melting the highest stages in the months of the spring. In the summer the incresed evaporation. the water consumption of the plants and not least

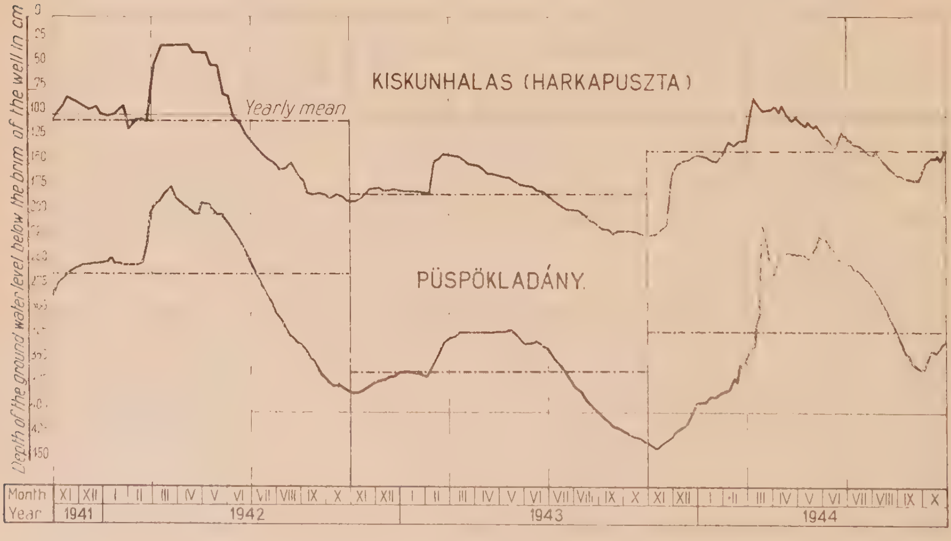


Fig. 2.

the general aridity steadily lowers the water table which culminates in the fall in the lowest position. The rains at the beginning of the fall and winter raise again to some extent the water table, and depending on the inclemency or the mildness of the winter, turns in the melting period in a comparatively higher level. Figure 2. represents the variations of the ground water wells at Kiskunhalas-Harkapuszta and Püspôkladány for the hydrologic years of 1942, 1943 and 1944. (Location of the wells may be found on Figure 8). The well at Püspôkladány is characteristic of the even territory of the trans-Tisza while that at Kiskunhalas of the territory between the Danube and the Tisza. It is most peculiar that at the trans-Tisza well the water table was considerably lower below the ground level than that of the well of the territory between the Danube and the Tisza. This circumstance is generally characteristic of the ground water conditions of the two regions. In the case of the two examined wells, the spring maximums were by 140 cm to 180 cm, the fall minimums by 190 cm to 210 cm lower at Püspôkladány than those at Kiskunhalas. Among the above three years, 1942 was very rainy and cool, 1943 was exceptionnaly poor of precipitation and hot, while 1944 could be considered a year of average climatic conditions. This circumstance is decisively manifest in the ground water stages of the respective years. In rainy years the minimum anteceded by the maximum at Kiskunhalas shows a stage of less than 160 cm while that at Püspôkladány shows less than 200 cm. As a whole similar stage differences prevail between annual minimum of the arid year and the maximum of the following average year. In an arid year the differencies between the maximum of the spring and the minimum of the fall are considerably smaller than the foregoing values. Thus, for instance, they are at Kiskunhalas about 80 cm, and at Püspôkladány about 120 cm. The annual mean stage in a rainy year is at Kiskunhalas by 40 to 60 cm,and at Püspôkladány by 60 to 90 cm higher than in the average years. While a dry year compared with an average year shows a difference at Kiskunhalas of 20 tot 40 cm and at Püspôkladány of 40 to 80 cm. As a result it appears that the ground water stages at Kiskunhalas and at Püspökladány with their spring maximum and their fall minimums show that the ground water conditions of the Alföld are continental. On both wells the variations of the ground water stages, considering the division of *Ijjász,* are characteristic of the so-called ground water type *fed by infiltration.* It must be mentioned that in viewing the ground water conditions on the Alföld besides the *average* and *deep normal types* the forest *type,* the type *fed by rainfall, the type developing from hydrostatic pressure,* and the *indisturbed type* mus also be taken into consideration.

Figure 3 represents the 10 years duration of the examined ground water wells. The ten years taken into consideration, extending from November 1, 1937 to October 31, 1947. considered as a decade of average hydrologic conditions, since the rainy and arid years are about equally represented. From the figures it may be perceived

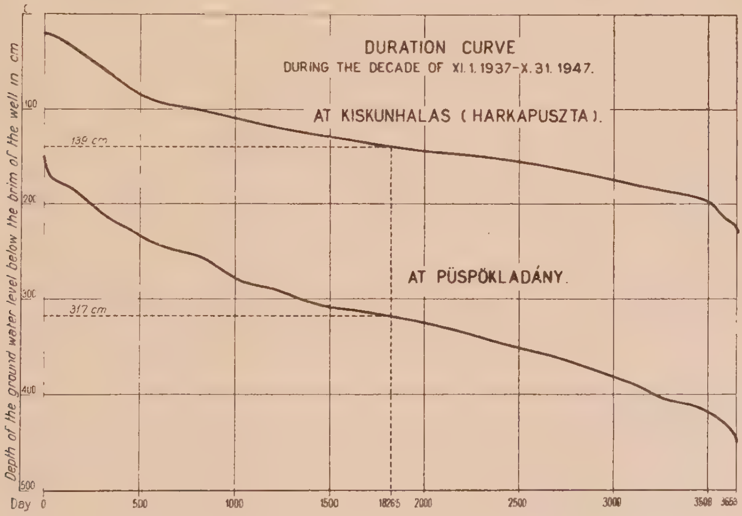


Fig. 3.

as to how many times occurred higher or lower stage within the ten years compared with a given stage. The average stage, that is the 185.5 days duration stage referring to one year, was at Kiskunhalas 139 cm and at Püspökladány 317 cm.

Figure 4 shows the annual mean stages for 14 years at Kiskunhalas and at Püspökladány. This figure discloses plainly the identical variation of the arid-hot, the average and the rainy cool years on both wells. The variations recorded in the order of time on the two wells of different trait, located in different districts verify also the hydrographic unity of the ground water conditions of the Alföld.

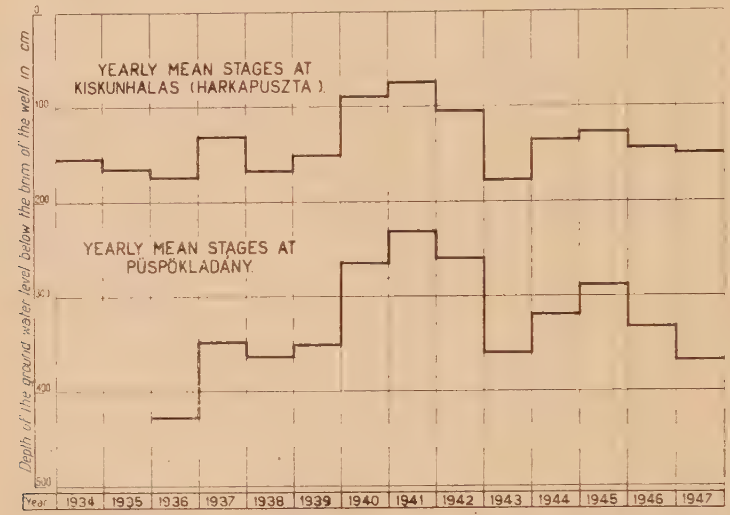


Fig. 4.

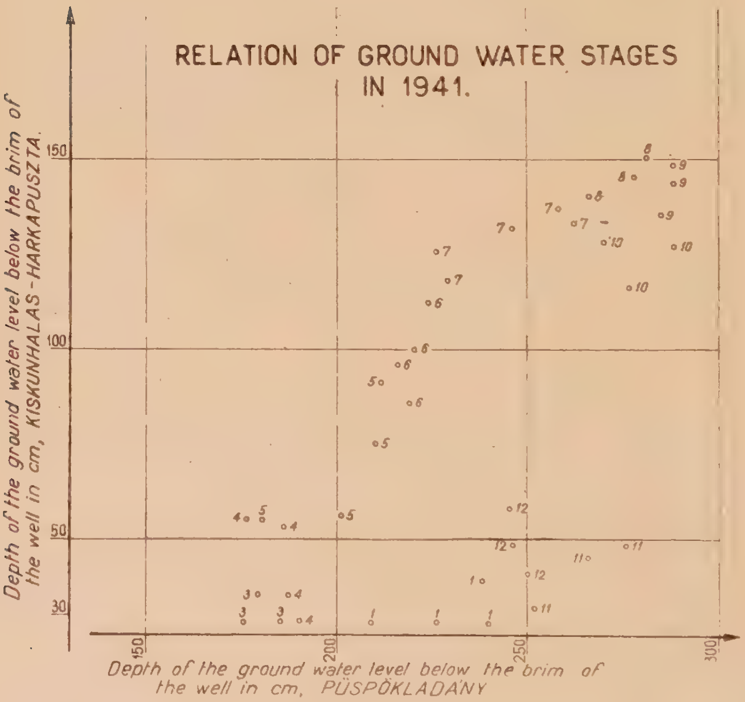


Fig. 5.

In accordance with the foregoing the question may arise : is there any demonstrable relation between the stages of the wells located in different districts of the Alföld ? In adjacent, closely together located wells by all means there can be found more or less distinct relationship between the ground water stages since the local conditions influencing in the first place the water tables can be considered identical. In the stage of the wells located of some distance from each other we hardly expect, however, any relationship. Should, however, by any chance such relationship appear it would prove only the close correlation of the stages with certain elements, with elements identical or at least similar at both wells.

In this respect we have made our examination on the two wells of the Alföld located far apart from each other, i.e. on that at Kiskunhalas and the one at Körösladány. According to *figure* 5 we have tried in 1941 to determine the relationship of the stages of the above mentioned wells. Each point signifies stages recorded on the same day. The Arabic numerals next to each point signify the month of the recording from which the characteristics of the stages may be deduced. Figure 2 means namely in general, that in the 11th, 12th, 1st and 2nd months there were rising stages while in the 3rd, 4th and 5th, or in the preceding or in the following months there were culminating, and in the 6th, 7th, 8th, 9th and 10th months there were falling stages. The year of 1941 was exceptionally rainy and cool, and according to the figure 5 there appears no relationship whatsoever between the stages.

Figure 6 also represents the relationhsip between the stages at Kiskunhalas and Püspökladány. The year of 1944 may be considered as a year of average weather condition. Although there is a wide scatter with respect to the falling stages nevertheless some relationship may be observed.

Figure 7 finally shows the stages at Kiskunhalas and Püspökladány in the very arid year of 1947. According to the figure the relationship of the stages of the two wells is represented by a loop like curve. One stem of the curve shows the rising, the other the falling stages. The end of the curve represents the culminating, and of the preceding and the following stages showing that the culmination occurred sooner at Körösladány than at Kiskunhalas. In recording of the minimums no displacement occurred, and the stages at the lower culmination are represented also by the stem or its extension of the « falling » curve.

On the basis of the three figures it may be said that in rainycool years no relationship can be found between the stages, in average years only slight or partial relationship is apparent, while in aridhot years the relationship is proven. The foregoing brief approximate comparison shows therefore that the ground water stages are in closest correlation with the climatic conditions. In the case of extreme precipitation this correlation is not manifest, just because the decisive, but disturbing influence of the precipitation. Naturally the examination does not assert that there exists no correlation between the stages and the precipitation. Such correlation evidently exists but its expression is very complicated since many other factors would have to be considered at the examination. The foregoing was expounded and expressed by scientific means by Endre Németh professor at Budapest to be read in his paper « Study of the correlations existing between the variation of depths and the temperatures of the ground water. »

Finally we wish to submit for a more detailed illustration of the ground water conditions of the Nagy Magyar Alföld (Great Hungarian Lowland) situation plan with hydroisohipsas and depths below the ground of a year with a rainy-cool spring and of a year with a comparatively arid-hot fall. The situation plans with hydroisohipsas were made on the basis of mean stages of each month which were taken into account for each well separately. *Figure* 8 shows the hydroisohipsas drawn up on the basis of the mean stages of the month April 1941 with respect to the elevation above the Adria. *Figure* 9 also shows the mean stages of the same month refering to the local ground levels. From the latter figure the depth is shown of the water table below the ground level in the examined rainy-cool month at the diverse locations of the Alföld.

From figure 8 the direction of the streaming of the water table of the Alföld can be perceived. Superficially following the line of the water divide between the two great rivers, the Danube and the Tisza, the highest line of the water table runs from north to south, and is sloping to the right and left to the Danube and the Tisza respectively. In this locations the slope is frequently 20 to 25 ⁰⁰/₀.

The slope appears somewhat less in the region of Nyiregyháza (15 to 20 ⁰⁰/₀), which, however, decreases in the south-western direction towards the Tisza so that in the heart of the trans-Tisza region the slope of the ground water table is 1,5 to

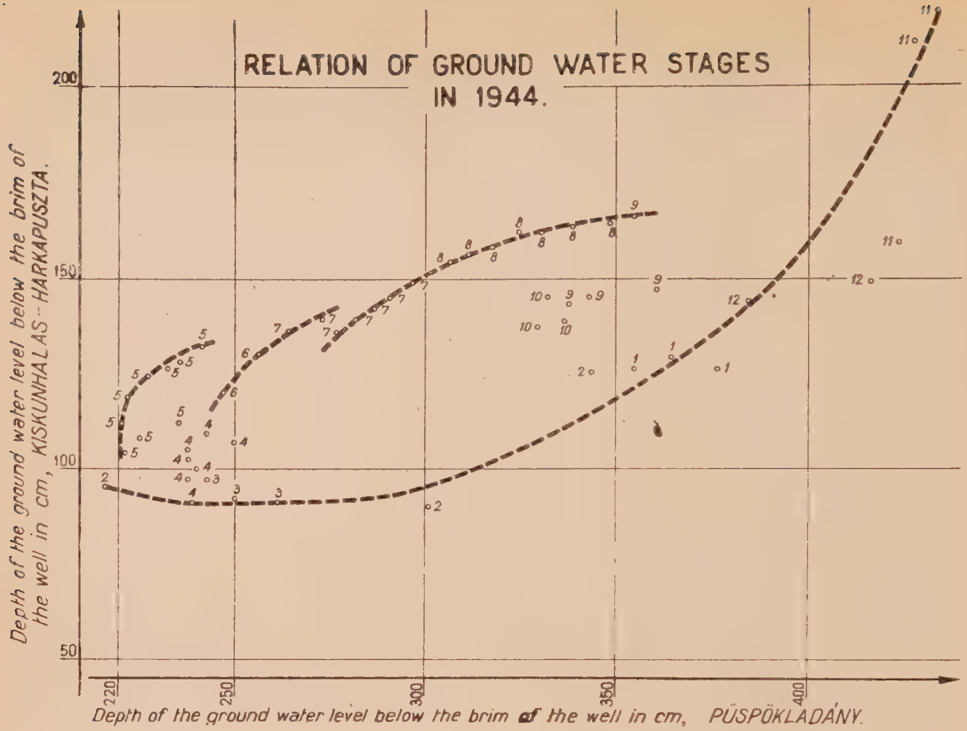

RELATION OF GROUND WATER STAGES
IN 1944.
Depth of the ground water level below the brim of the well in cm, KISKUNHALAS-HARKAPUSZTA.
200
150
100
50
220
250
300
350
400
Depth of the ground water level below the brim of the well in cm, PÜSPÖKLADÁNY.

Fig. 6.

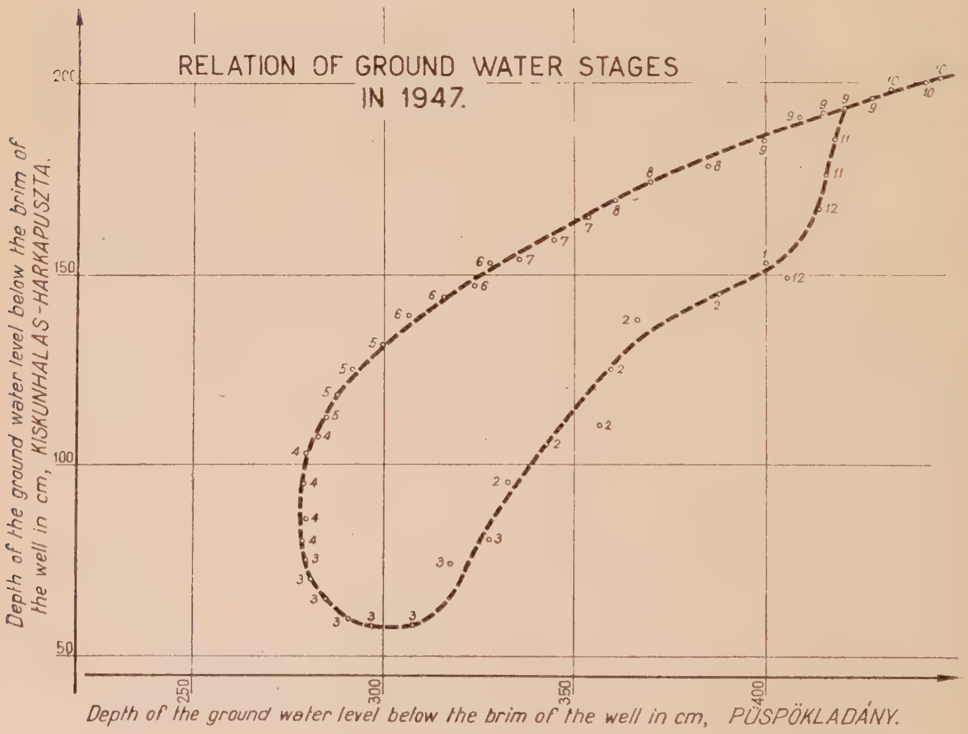

RELATION OF GROUND WATER STAGES
IN 1947.
Depth of the ground water level below the brim of the well in cm, KISKUNHALAS-HARKAPUSZTA.
200
150
100
50
250
300
350
400
Depth of the ground water level below the brim of the well in cm, PÜSPÖKLADÁNY.

Fig. 7.

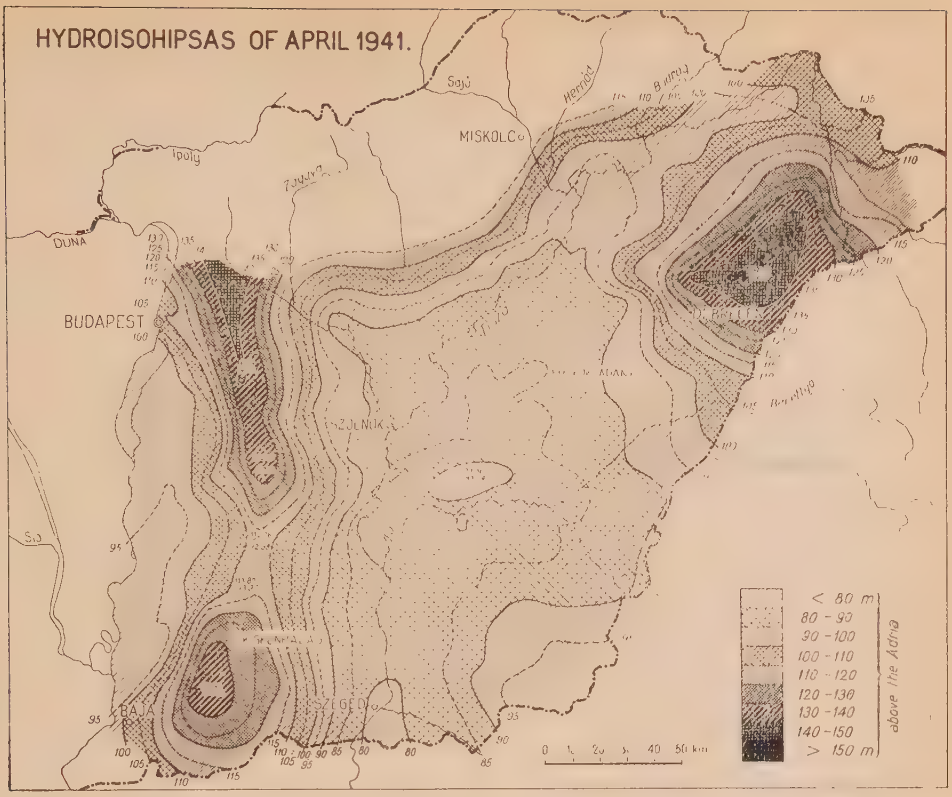


FIG. 8.

2.5 ⁰⁰/₀ only. At some singular wells we have also indicated by numbers the stage of the ground water above the Adria. Below the numerals we have indicated with italics at the same wells the elevation of the ground level.

Figure 9 represents, as we have already mentioned in the foregoing, the water table below the ground level as observed in the month of April 1941. The lines show a great variation; which is mainly dependent on the variations of the soil composition. Aside of the ground water depth of small extent — 5 to 6 m — in the north-east the ground water table in the region of the Danube — Tisza was in general considerably higher than in the trans-Tisza region. In a few places the ground water table was found to be higher than the ground level on account of inundation. According to the figure the ground water can be found between the Danube and the Tisza 1 to 2 m, in the trans-Tisza region 2 to 5 m below the ground level in a year vith maximal precipitation.

Figure 10 shows the hydroisohipsas compiled on the basis of the mean stages of October 1943. The fall month of this aridhot year shows extremely low stages. It is of peculiar phenomenon that the sloping of the ground water and the direction of streaming of this low stage is similar to the conditions apparent in the high ground water as stated in figure 8. The slopes are somewhat less than at high stages but there can be found also here as in the region between the Danube and the Tisza and in the region of Nyiregyhâza in the trans-Tisza district slopes of 23 ⁰⁰/₀. In the trough of the trans-Tisza district, however, slopes of less than 1 ⁰⁰/₀ may be found also. On a few singular locations of figure 10 we have especially marked also the ground water table and the local ground level above the Adria.

Similar to figure 9 in *figure* 11 we have shown the water table of October 1943 below the ground level. In the case of such extremely low stages the water table below the ground level isvery capricious also in the present instance. The comparison of the high stage with the low stage naturally is not possible any longer in this

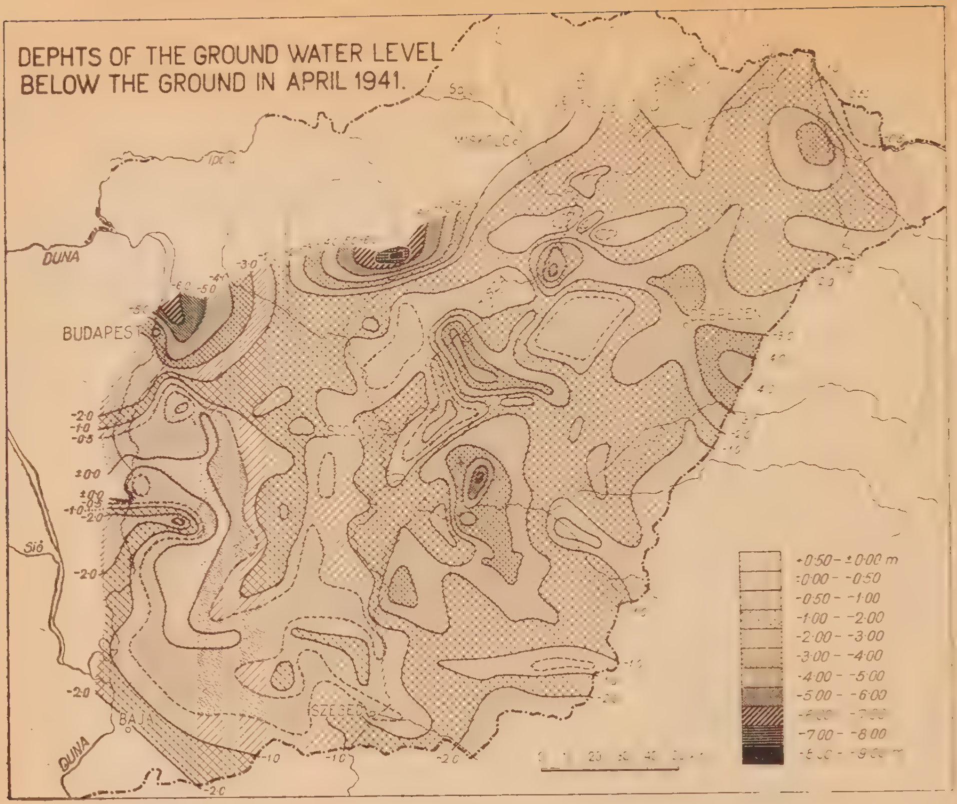

DEPHTS OF THE GROUND WATER LEVEL BELOW THE GROUND IN APRIL 1941.
DUNA
BUDAPEST
SZEGED
BAJA
Sió
-2·0
-1·0
-0·5
±0·0
-3·0
-5·0
-6·0
-10
-2·0
+0·50 – ±0·00 m
±0·00 – −0·50
−0·50 – −1·00
−1·00 – −2·00
−2·00 – −3·00
−3·00 – −4·00
−4·00 – −5·00
−5·00 – −6·00
−7·00 – −8·00
0 10 20 30 40 50

Fig. 9.

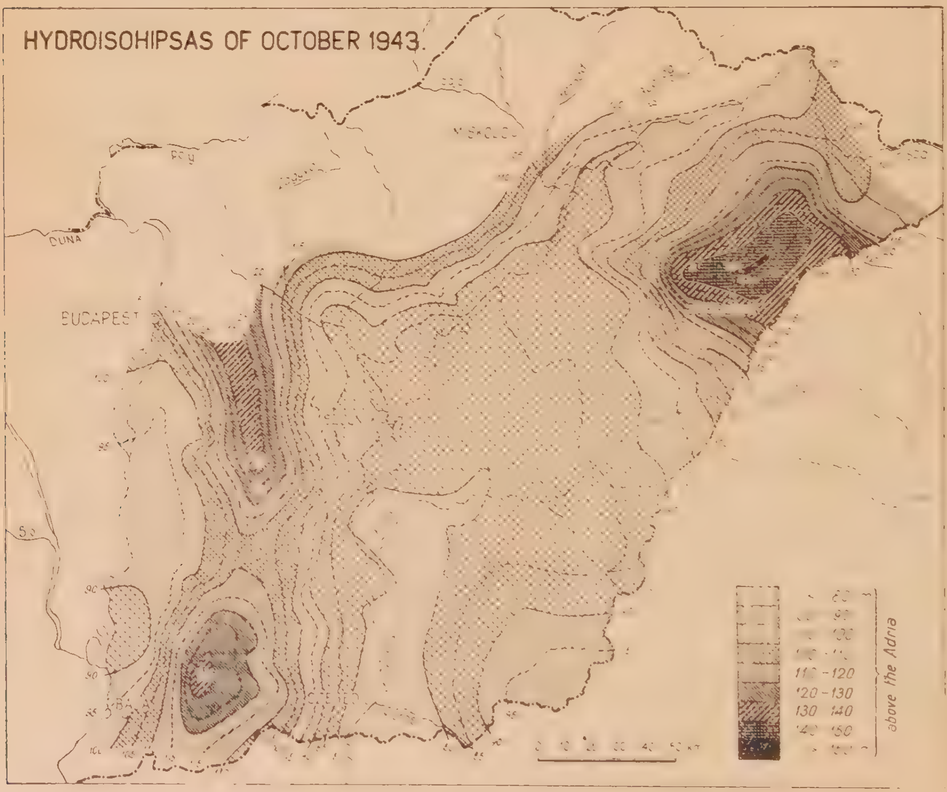

HYDROISOHIPSAS OF OCTOBER 1943.
DUNA
BUDAPEST
MISKOLC
BAJA
90
110 – 120
120 – 130
130 – 140
140 – 150
above the Adria
0 10 20 30 40 50 km

Fig. 10.

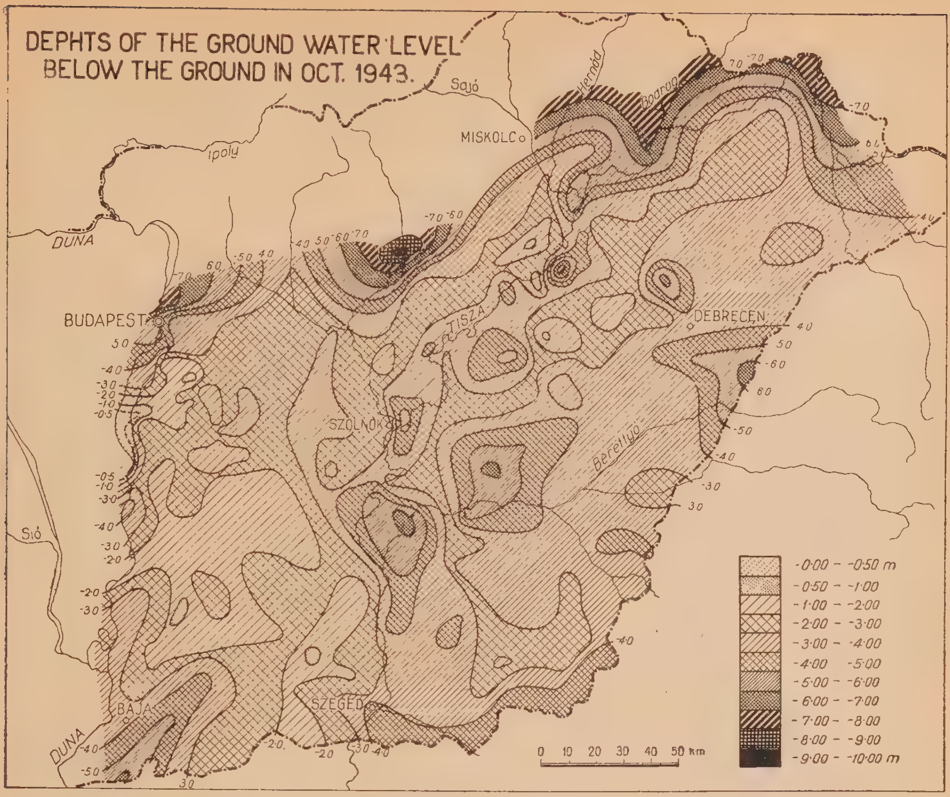


FIG. 11.

case since in both instances the composition of the soil exercice a great influence in forming of the ground water table. There can be found here in the northern part between the Danube and the Tisza below the ground level of about 10 m water tables, moreover these are even in the region of the Bodrog river depths of 7 to 8 m. In the region between the Danube and the Tisza the ground water tables are located in general at 2 to 5 m while in the trans-Tisza region at 3 to 5 m below the ground level.

The exhibited hydroisopsas and the maps showing the water table below the ground level were made on the basis of the respective wells, after the necessary interpolations were finished, and they characterize in general only in their main feature the ground water conditions of the Alföld. There might occur differences of some local importance but nevertheless we assume that these maps are the most dependable among those which have been published so far for the information on ground water conditions; on the basis of a well network of this extent no hydroisopsas were published before about this region.

4)

GENERAL REPORT ON GROUND-WATER INVESTIGATIONS IN THE UNION OF SOUTH AFRICA SINCE 1939

by

Louis T. NEL
Director-Geological Survey

Since that date our work has progressed slowly, at first due to the war and latterly due to shortage of man-power which has caused considerable wastage among the staff. During this period the Geological Survey of the Union of South Africa provided a militarised section which was mainly concerned with locating sites for drilling for water for the troops in East Africa and the Middle East.

At home the site locations for boreholes for farmers and State undertakings continued with success though the terrain is becoming more difficult. More use is being made of the electrical resistivity apparatus, and the vertical magnetic variometer as an aid to tracing suitable water-bearing conditions underground.

The following papers were published :

1. Scientific Methods of Water finding. Presidential Address to the Geological Society of South Africa, 1943, by H. F. Frommurze D.Sc.
2. Basins of Decomposition in Igneous Rocks : Their Importance as underground Water Reservoirs and their Location by the Electrical Resistivity Method by J. F Enslin M.Sc. Transactions of the Geological Society of South Africa Vol. XLVI 1943.
3. Notes on some Magnetometric Surveys with Special Reference to Anomalies on Karroo Dolerites by B. D. Maree M.Sc. Transactions of the Geological Society of South Africa Vol. XLVI 1943.
4. A Geochemical Survey of the Underground Water Supplies of the Union of South Africa, with particular reference to their utilization in power production and industry. Geological Survey Memoir No. 41, 1946.

Work on Infiltration (soil-absorption) : Factors that determine the infiltration part of precipitation, including the effect of plants, land configuration etc. and the magnitude of these influences have been carried out fairly comprehensively in very limited areas by the Research Section of the Forestry Department at Jonkershoek and Cathedral Peak.

Occasional investigations which may include all phases of the water-cycle have been made on a modest scale and conducted over short periods, only, by the Pretoria and Witwatersrand Universities.

At present a comprehensive scheme of hydrological research to cover the whole area of the Union of South Africa is under discussion by the South African Council for Scientific and Industrial Research and the various Government Departments and other Institutions and Agencies which have studied those special phases of the cycle required for their particular work.

5) CASES OF EXTREME ABUNDANCE OF GROUND WATERS IN CZECHOSLOVAK REPUBLIC

Ota HYNIE
(State Geological Institute Prague) :

(With a Hydrological Map and Profiles)

Near the northeastern border of the Palaeozoic mountain system, which emerges in the heart of Europe from under the younger formations and is known as the Bohemian Mass, occurs a relatively thin cover of sediments of the Cretaceous formation extending over an area of about 16.000 square kilometres. This area comprises northern, northeastern and eastern Bohemia; it continues from Bohemia on the one hand to the NW with a tongue along the Elbe into Germany as far as Dresden, and on the other hand to the SE into Moravia across the Svitava River as far as Blansko near Brno. This Cretaceous region has locally a petrographic development of its beds and tectonics to the origin of *exceptionnally abundant reservoirs of ground waters.* These manifest themselves as very abundant overflow springs, and in the great yield of the wells by which the Cretaceous ground waters are exploited for water-conduits of drinking water. The hydrological conditions of some reservoirs of ground water are well known and allow the *computation in figures of the specific outflow of the ground water, which for certain regional types are generally valid.*

The Bohemian Cretacious series is of Upper Cretaceous age and includes the Cenomanian, Turonian and Lower Senonian. The beds were deposited in a shallow sea, in a geosyncline with its axis running from NW to SE, parallel to the tectonic northeastern border of the Bohemian Mass.

From the geological map and profiles it is evident that the extensive, but shallow Cretaceous basin with its maximum depth of about 500 m. is disturbed mainly by faults parallel to the NE border of the Bohemian Mass, i.e. by faults parallel to the axis of the basin. Along these faults the border blocks subsided compared with the centre of the basin, so that the greatest thickness of the Cretaceous beds is often found in the limbs of the syncline and not in its centre. On the lower course of the Bohemian Elbe, approximately between Usti n. L. and Českà Lîpa, where the Cretacious series is traversed by many Tertiary eruptions of a basaltic type, the Cretaceous series is traversed by normal faults of a second system, of a SW-NE direction, parallel to the Metalliferous Mountains on the NW border of Bohemia. The faults of this second system form in the Cretaceous series a deep transverse rift.

The faults of these two systems were first formed in the Mesozoic or already during the Carboniferous Hercynian folding, but the chief movements along them, by which the Cretaceous series was dislocated, did not take place until the Tertiary period. A dense network of fissures runs parallel to the faults of the two directions in the Cretaceous series; to these are added the youngest fissures of a third system in the Jizera direction (N-S with a slight twist into the direction NE-SW); although these youngest fissures occur the most rarely, yet their openness makes them the most important ones for the circulation of the ground waters.

In its eastern half the main Cretaceous basin is accompanied on its northeastern and southwestern side by separate basins, the largest of which is the separate basin in the neighbourhood of Police n. Metujî, as well as by narrow tectonic rifts.

Petrographically the beds of the Bohemian Cretaceous series are of three types : On the borders of the Cretaceous sea and only in the lowest, overlapping Cenomanian group in the whole Cretaceous area of sedimentation *sandstones* were formed of the sandy sediments of river deltas; farther from the mouths of the rivers were formed *marly sandstones and sandy marls* and within the Cretaceous sea *marls and clays.* In the course of sedimentation in the shallow Cretaceous sea the regime of the rivers changed, which carried material into the sea, and also the depth of the sea changed. This caused the facial variety of the Cretaceous beds in their areal extent and in a profile across the beds.

The *Cretaceous sandstones* are pure quartzitic sandstones of predominantly medium grain. They are insufficiently cemented, crumbly and disintegrate easily. They are arranged in thick banks of square disintegration, and are therefore called Quader sandstones. The volume of the pores amounts to between approximately

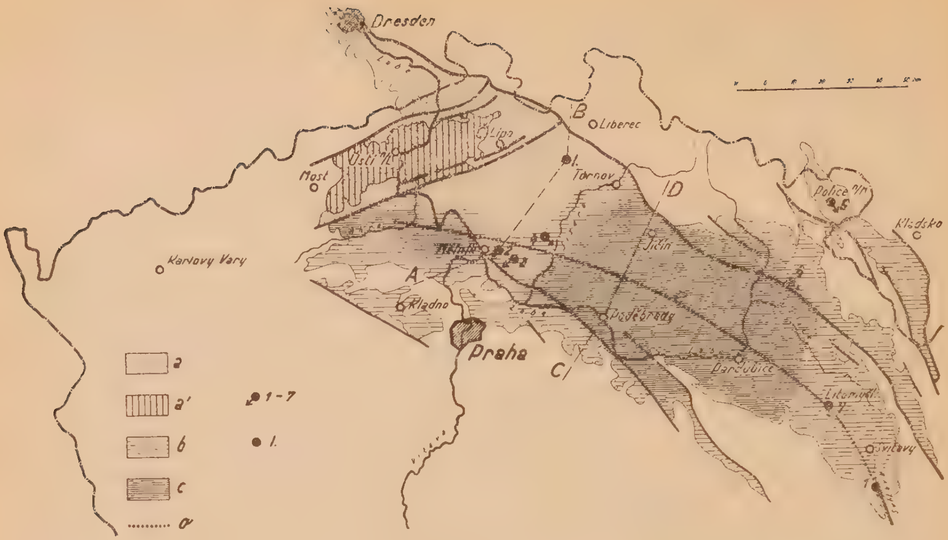


FIG. 1. — *Hydrogeological Map of the Cretaceous Formation of Bohemia*
a = *Quader sandstones,* highly permeable
a' = the same under an Upper Tertiary cover.
b = marly sandstones and sandy marls, predominantly with fissure ground water.
c = impermeable *marls and clays.*
o = Axis of the main Cretaceous basin.
1- 7 = spring groups, with indication of the yield in l/sec.
(1 = 750-1000 l/sec, 2 = 300 l/sec, 3 = 250 l/sec, 4 = more than 100 l/sec, 5 = 150-170 l/sec., 6 = 130 l/sec., 7 = more than 100 l/sec.).
I = artesian bore with a yield of 100 l/sec.

15 % and 30 % of the total volume of the rock. Besides pore water the Quader sandstones contain, however, also fissure ground water, as they are intensively fissured. Their fissures are often open, but are usually filled with sandstone fragments with a good filtration effect.

Marly sandstones and sandy marls, though these have a considerable pore volume up to over 10 % of the total volume of the rock, they have very fine pores, which usually bind the water by capillary forces. On the other hand they are intensively fissured. Their ground water is typical fissure water. More copious springs and the greater yield of catchment wells are exclusively due to fissures both in beds of this type as well as in the Quader sandstones.

Marls and clays, even when finely arenaceous, have only in exceptional cases a small contents of fissure ground water. Usually the fissures are closed in them.

By the alternation of Cretaceous beds of these three types *several horizons of ground water* are formed in the Cretaceous series of Bohemia. When permeable beds rise to the surface, the uppermost ground water horizon has a free level; the deeper horizons have artesian, confined water.

The lowest ground water horizon in the Cenomanian sandstones extends almost throughout the whole of the Cretaceous formation continuously. About one third of the Cretaceous cover, its central part, which is situated between the Elbe and the Jizera (see map and profile C-D), has besides this basal Cenomanian sandstone the whole sequence of beds composed of impermeable marls; the outer sandstone region and the region of marly sandstones and sandy marls beyond the border faults of that central crea of the Cretaceous cover of Bohemia were denuded. Only the eastern and especially the northwestern part of the Cretaceous cover have permeable beds developed also above the Cenomanian, especially in the middle Turonian (in the zone with Inoceramus labiatus and Inoceramus Larmacki) and in the Upper Turonian (in the zone with Inoceramus Schlönbachi). These permeable beds form also several horizons of artesian water above each other. Their artesian pressure is more frequently caused by a different degree of permeability of conformably deposited beds than by the alternation of permeable and impermeable beds.

The different ground water horizons combine over large areas to form more or less uniform reservoirs of ground water; the height of the column of beds saturated with water may even considerably exceed 100 m. When such a favourable stratigraphic development of the beds is combined with favourable conditions for the infiltration of atmospheric water and especially with favourable local tectonic conditions, ground water *reservoirs of an enormous yield are formed.* Favourable tectonic conditions are given especially by separate basins and the uplift of the ground water flows on tectonic faults or also on basaltic walls filling the faults.

The overflow springs of the Cretaceous ground water are numerous and copious There are in Bohemia and Moravia about 115 concentrated springs with a minimum yield of more thaf 5 l. /sec. Some 25 springs have over 25 l. /sec., and 3 springs have a minimum yield of over 100 l. /sec., two of them of over 250 l.–sec. Over 10 springs have a maximum yield of over 100 l. /sec. (according to Ing. F. Podvolecky). As a rule several abundant springs occur side by side, so that the yield of such *groups of springs* may attain even several hundred l. /sec.

The Cretaceous springs, especially the springs of great yield, are very constant. Usually the ratio of the minimum discharge to the maximum discharge is about 1 : 2, at the most 1 : 3; the natural artesian springs using for their rise from the artesian basins to the surface tectonic fractures, especially the intersection of two tectonic faults, have a discharge varying only within quite narrow limits, they are practically constant. They betray themselves also by their temperature, which is a few degrees higher than the mean annual temperature at the site of the spring.

The most abundant group of springs is situated in the southeastern corner of the Cretaceous formation, in the valley of the river Svitava in Moravia SE of the town of the same name (ad 1 in the map appended). Owing to the inclination of the length axis of the principal Cretaceous basin to the SE combined with the simultaneous wedging out of the basin 750-1000 l. /sec. are here discharged within a relatively small area in six concentrated springs; two of these springs have a minimum yield of 250 l. /sec. (one of them, the spring at Banîn, is captured for the water supply of the town of Brno; the extension of this water supply by the capturing of further springs is planned).

The second group of springs, as far as yield is concerned, is situated east of Melnîk (2). A minimum of 300 l. /sec. overflows in a few springs close to each other from the concentration of the *ground water in the brachysyncline* (see profile A-B) into the Pšovka valley. Of these the concentrated spring at Mělnické Vrutice has alone a discharge oscillating between 120 and 260 l. /sec. Not far from this group of springs another 250 l. /sec. are discharged in the valley of the Košatec stream from three pairs of neighbouring fissure springs (3). and a little farther to the NE in the valley of the Strenice stream (4) more than 100 l. /sec. from overflow springs of two horizons of ground water are discharged. (The utilisation of all three groups of springs is planned for the water supply of the Municipality of Prague).

From the separate Cretaceous basin at Police n. Metujî about 150-170 *l. /sec.* are discharged in five places not too far from each other (5). They are overflow springs and natural artesian springs.

At České Meziřîčî NE of Hrádec Králové (6) three springs situated close together have a constant yield of about 130 l. /sec. They are natural artesian springs from tectonic fissures.

In the neighbourhood of Litomyšl (7) is a group of springs, predominantly natural artesian springs, of a total minimum yield of over 100 l. /sec.

Besides being utilised in very numerous shallower wells, the ground waters of the Cretaceous formation are utilised for *more than* 300 *bored artesian wells* with depths of from some tens of metres to 400-500 m. The yield of these wells is also large. A yield of 10 l. /sec.—30 l. /sec. is not rare. Only the deep artesian wells (100-400 m.) have a smaller yield (1 l. /sec.—5 l. /sec.); they are bored in the region of a marly facies of the Cretaceous beds between the Elbe and the Jizera to the basal artesian horizon of the Cenomanian sandstones, to which infiltration of water from the surface is difficult.

The most copious well is the bored well NW of Turnow (I) in the Zâbrtka Valley at the village Křida, intended for the water supply of the town of Liberec. In lowering the water level 7 m. under the surface 100 l. /sec. were drawn from this well, 222 m. deep, *but only after the torpedoing of the bore,* by which the originally small yield was increased by the inflow of water from fissures opened by the torpedoing of the bore. The well lies at a basaltic wall uplifting the artesian water of two hozizons, of which the lower horizon supplies the predominant part of the inflow (see profile A-B).

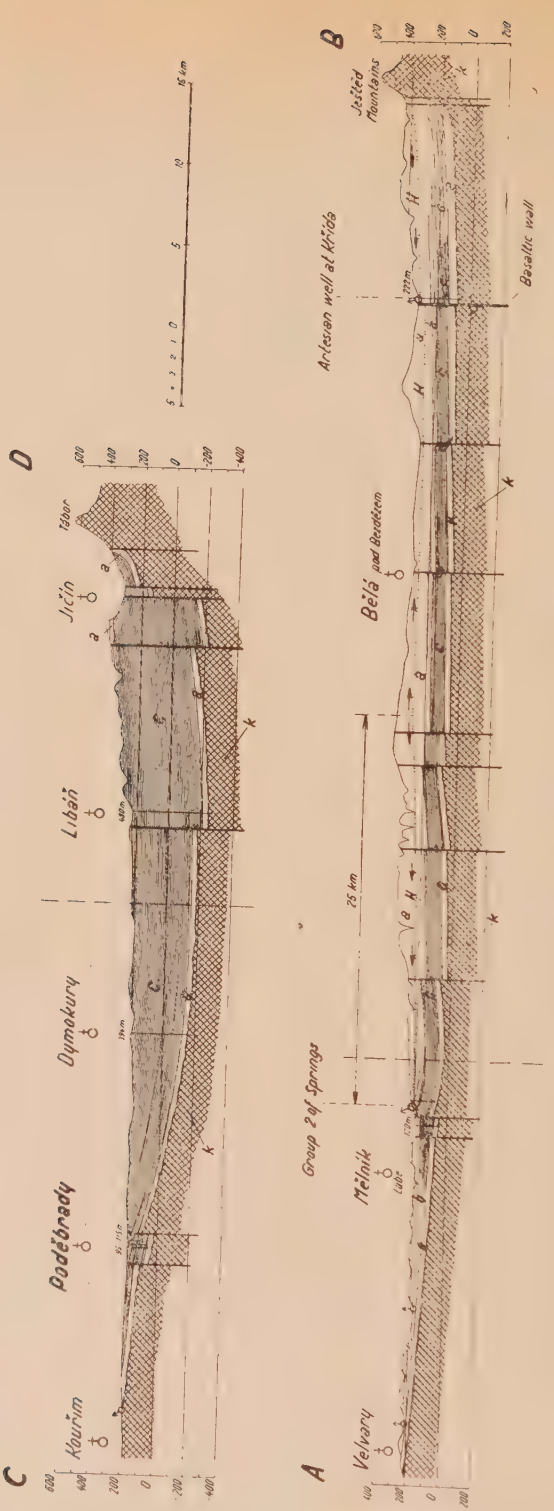


FIG. 2. — *Hydrological profiles across the Cretaceous formation of Bohemia.* (vertical scale ten times exaggerated)

■ = Quader sandstones
b = marly sandstones and sandy marls
c = marls and clays
d = base of the Cretaceous formation
H = Level of the underground water (free)

In the exploratory borings undertaken in the areas of copious springs for the purpose of their utilization as supplies of drinking water it was found that the water of these springs can for the greater part be more advantageously captured by capturing the ground water flows (by systems of suitably chosen wells) supplying the springs than by the direct capturing of these springs themselves.

All in all there are in Bohemia and Moravia *six* regions with copious reservoirs of Cretaceous ground waters, all of which do not, however, betray themselves by exceptionally copious springs. After completely supplying the local population of these regions with drinking water, there remains in them a *surplus of about* 4000 *l. /sec. of ground waters economically capable of utilisation for the supply of drinking water* of larger towns, and especially of densely populated manufacturing regions situated in territory poor in ground water. Some of the regions rich in Cretaceous ground water make possible a further enrichment of their reservoirs of ground water by infiltration of water from rivers and surface reservoirs (*formation of artificial ground water*).

Their physical, chemical and bacterial purity make the Cretaceous ground waters apart from a few isolated exceptions, excellent supplies of drinking water of medium to great hardness.

* * *

Exact knowledge of the hydrological conditions of some regions rich in Cretaceous ground water makes possible the determination of the *specific outflows of ground water* from areas characterised by a certain geological composition, landscape character, precipitations and other circumstances influencing the formation and movement of the ground water. I give in the following the principal results.

1) *Regions of Quader sandstones* : Uniform composition with a total thickness of the sandstone series, always intensively fissured, of several tens of metres. Arrangement in very gently sloping basins or tables with a very gentle inclination (1—5°), on a practically completely impermeable base. The level of the ground water is free, the thickness of the beds saturated with ground water is of up to a few tens of metres. The ground water is drained by overflows (springs and diffuse seepages) at the border of the blocks of Quader sandstones, which in the hill region form the summits of hills dominating their neighbourhood, and which cover an areas of 1-4 sq. km. The surface of the Quader sandstones is rugged, and consists of a maze of sandstone rocks. In the depressions between the rock formations is sandy waste. Scanty pine covering (the surrounding country mainly forested). Under these the most favourable conditions of infiltration imaginable the *specific outflow of the ground waters* directly measured in the overflows of their ground waters amounts to 6 l. /sec. /sq. km. with a small amplitude (Drabovna near Turnov) at an altitude of 450 m. above sea level and with an average rainfall of 850 m. /m.; at an altitude of 820 m. above sea level and with an average rainfall of 900 m. /m. it oscillates between 7 and 8 l. /sec. /sq. km. (Bore near Police n. Metují). In the first case about 22,5 % of the total rainfall measured permeate to the ground water reservoir, in the second case 25-28,5 %.

2) *Regions composed of tectonically disturbed marly sandstones and sandy marls* : Arranged in gently sloping basins or tables with a very gentle inclination. Thickness up to more than 100 m. Gentle hill country and plateaux with sharply incised erosive valleys, covered with loams, predominantly loess loams, cultivated, and only for a small part forested. Large areas of the country, of several hundreds of square kilometres, with outflows ascertained mostly by measuring the minimum flows through stream cross sections. With an average rainfall of 500 m. /m.—600 m. /m. the *minimum specific outflow is around* 2 *l. /sec. /sq. km.* according to measurements in different regions and rather below this. This corresponds to an infiltration of 10-13 % of the total rainfall measured.

3) In a region of mixed geological composition (in the Pšocka basin at Mělnik) the specific outflow is determined by the direct measuring of springs and diffuse seepages of ground water in a well-known geological river basin of an area of about 200 sq. km. The region is one fourth Quader sandstones, three fourths marly sandstones and sandy marls, with the characteristic relief of both types as in the cases described above. With an average rainfall of 520-600 m, /m. the specific outflow is 2,3 l. /sec. /sq. km. minimum outflow, 4,5 l. /sec. /sq. km. maximum outflow.

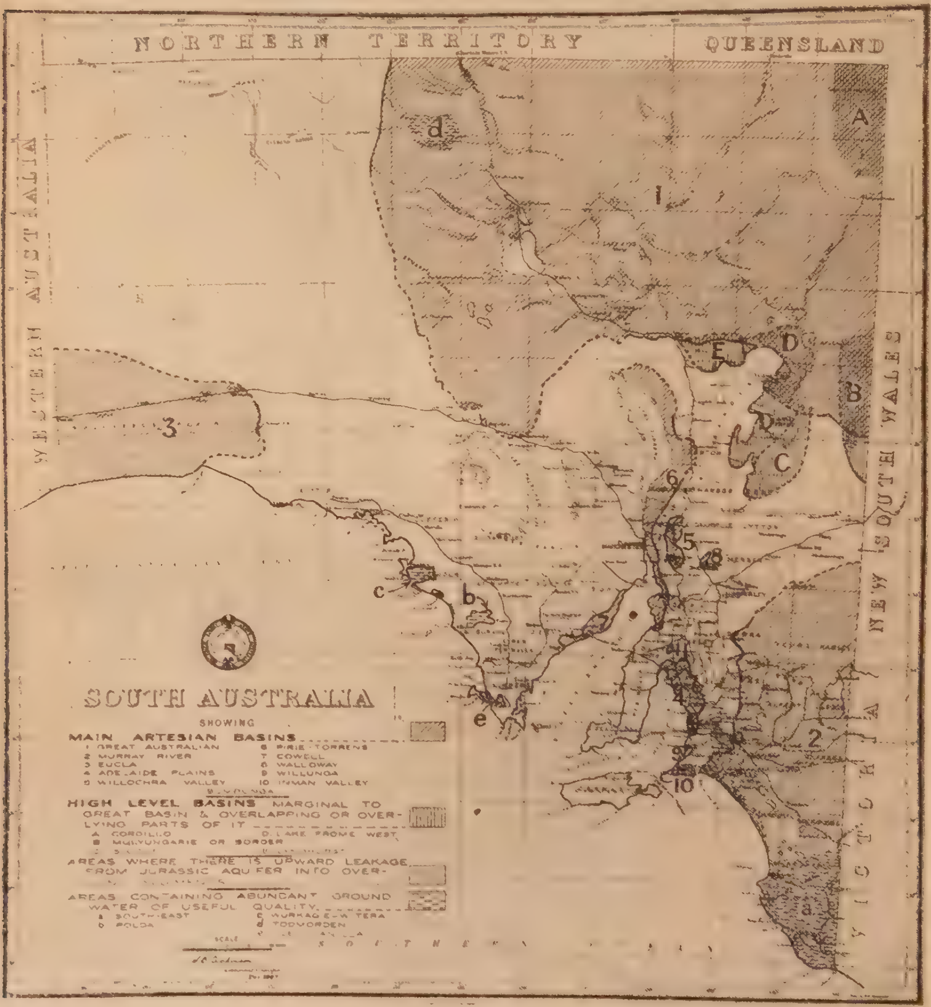
Other regions in which the minimum specific outflow was determined by measuing the minimum flow in the streams at the end of long rainless periods, have more complicated conditions, especially unequal composition, so that the figures obtained have no general validity.

6) UNDERGROUND WATER IN SOUTH AUSTRALIA

by

S. B. DICKINSON

When judged by any standard, South Australia cannot but be regarded as a region in which conditions of aridity predominate. The areal extent of the State is 380,070 square miles, and 82,8 per cent of this total receives an annual rainfall of less than 10 inches. The excess of evaporation over rainfall is large, and much of South Australia exhibits a marked variability in the distribution of rainfall in all but the coastal areas. Hence it follows that very few streams discharge their waters into the sea. Many of these, rising in the highlands, disappear within a very short distance of their debouchure upon the piedmont plains. The River Murray alone, assisted by locks along its course and barrages at its mouth, remains a permanent stream of any magnitude.



Under such conditions the permanent occupation and full productivity of the land can be achieved and maintained only by the full development and conservation of the water resources of the State. There are many reservoirs the water from which is reticulated through a large proportion of the agricultural areas; part of the flow of the River Murray is pumped across the highlands through a pipeline that extends from Morgan to Whyalla on upper Eyre Peninsula; and it is probable that, at no very distant date, a further draught upon the Murray will be necessary to supply the expanding metropolitan area near Adelaide. Yet, with the utmost use being made of surface supplies, both natural and artificial, there is an ever present demand for water from the subterranean accumulations near the towns and throughout the farming districts and pastoral holdings. Several South Australian country towns are supplied from underground sources, and it has been found necessary on more than one occasion to force bore water into the Adelaide mains when drought conditions have caused the temporary depletion of the surface reservoirs.

The search for supplies of useful underground water is being conducted continuously, and it is possible already to indicate the artesian basins whence pressure water may be drawn, and also some of the sedimentary basins containing supplies of non-pressure water,as well as to state the conditions that must be satisfied before supplies of useful water can be obtained from the terrains in which dense rock occur.

CLASSIFICATION OF SOUTH AUSTRALIAN UNDERGROUND WATERS

The primary divisions into which the underground waters may be grouped are those determined by the pressure conditions governing the occurrence. On the one hand are the pressure or piestic waters found in the artesian basins, and on the other the ground or phreatic waters of which the upper limit is free and unconfined.

An artesian basin is taken to be the whole of an area within which pressure water exists (and from which artesian or sub-artesian water is obtained by boring), together with the area occupied by the ground or phreatic water contained in the upper and marginal unconfined portions of the water-bearing beds.

Ground or phreatic water does occur in defined sedimentary basins of considerable extent, but it is more commonly stored in openings of all kinds existing in dense rocks.

These basal considerations are taken into account in the following descriptive notes.

THE WATERS OF THE ARTESIAN BASINS

It will be seen, by reference to the accompanying map of the State, that the largest South Australian basins extend beyond its boundaries into adjoining territories. These shared basins are :

	Total Area sq. miles.	Areain South Aust. sq. miles.
The Great Australian Basin	578,000	120,000
The Murray River Basin	107,250	28,270
The Eucla Basin	68,500	16,500

The extent of the basins lying wholly within South Australia is shown in the following table :

	Area in sq. miles
Adelaide Plains	1,130
Willochra Valley	595
Pirie-Torrens	3,585
Cowell	635
Walloway	45
Willunga	50
Inman Valley	92
Bumbunga	40

In addition to these there are several high-level or marginal basins resting upon or bordering on the Great Australian Basin, as shown on the map.

The Great Australian Basin

a) *Stratigraphy and structure.*

This enormous basin which occupies about 22 per cent of the whole area of Australia is roughly triangular in plan, is filled with sediments the age of which ranges from Permian to Recent. The lowest beds, of Permian and Triassic age, outcrop on the eastern margin and probably underlie portions of the basin. They may act in part as intake beds, and may yield water from the deepest aquifers. The principal aquifer, however, is an exceedingly porous sandstone of Jurassic age, and of freshwater origin. It attains a thickness of about 6,000 ft. in Queensland, but South Australian boreholes have not penetrated deeply into the information. This waterbearing sandstone is overlain by a series of Cretaceous shales, the lowest of which are of marine origin and attain a maximum thickness of 2,433 ft. in northern South Australia. Conformably upon these marine shales rests a great thickness (3,695 ft. in the north-eastern corner of South Australia) of fresh water Upper Cretaceous shales with thin intercalated sandy beds and some lignitic matter. Upon these in turn there is a series of argillaceous and arenaceous sediments separated from the underlying shales by an unconformity. This series is believed to be of early Tertiary age. Finally the Recent windblown sands and fluviatile deposits of the Lake Eyre region constitute the uppermost beds of the basin.

There is evidence of some faulting and folding in portions of the basin, possibly connected with late Tertiary tectonic movements. In South Australia there are anticlinal structures, apparently due to differential vertical movements, in portions of the basin which are in line with the Flinders and the Peake and Denison Ranges near its southern margin.

The far south-western part of the basin in South Australia carries ground or phreatic water where the Jurassic sands extend over the Pre-Cambrian bedrock without a continuous cover of the Cretaceous shales. Useful water is obtained from the sands where there is absorption of rainfall shed from inliers of the bedrock or from outliers of Crecaceous shale.

b) *The source of the water; intakes and outlets.*

There is no doubt, in the minds of those who have studied the matter deeply, but that the water of the basin is of meteoric origin. The suggestion of a plutonic origin for any material fraction of the water is not regarded as tenable, and the only region of expiring or late vulcanism on the eastern margin has not proved to have made any contribution to the storage. Nor does it seem probable that any residual or connate water remains in the basin save perhaps in an isolated aquifer at Julia Creek in northern Queensland.

On the other hand the examination of the contours of the hydraulic or piestic surface shows that this imaginary surface rises in a regular way towards those areas in which absorption can and does take place. The major intake areas are located on the eastern margin in New South Wales and Queensland. In the latter State alone it is estimated that the intake beds cover 30,000 square miles and receive an average rainfall of 25 inches per annum,while in New South Wales their extent is estimated to be 9,500 square miles. In South Australia the main aquifer receives accessions of water, not from continuous outcrops of exposed porous sandy strata but from the incised valleys of the Finke and Alberga Rivers and Arckaringa, Stevenson and Hamilton Creeks which experience occasional heavy floods. It is possible also that a contribution is made by the sporadic flood waters of the Todd, Hale and Hay rivers which rise in the MacDonnell Ranges of the Northern Territory and terminate abruptly on the margin of the Arunta desert.

The only free outlet for the artesian water to the sea is that which takes place into the Gulf of Carpentaria. Geological structure precludes the possibility of any escape into the Murray River Basin or into the Southern Ocean to the South of the Nullarbor Plain. Nor does the conformation of the hydraulic surface give any support to such means of escape. There are, however, numerous natural outlets in the form of springs which have built up mounds or cones composed of calcareous sinter, sand and clay. These springs and the artificial boreholes together constitute the only known outlets for the water.

There are dissected remnants of immensely larger mound springs than any existing today at several places in South Australia, pointing to a much more abundant discharge of water under higher pressure than the present vents exhibit.

The outflow from the mound springs is lost by evaporation in the parched interior of the continent, as is the flow from the artificial boreholes.

c) *The Chemical character of the water.*

Analyses made over a series of many years have shown that the water in the main aquifer remains markedly constant in composition. Only two exceptions have been noted, and it is suspected that in these cases there may be contamination of the water by leakage from an upper aquifer by reason of corrosion or some other defect in the casing. It has been found that the water moving from an eastern source is characterised by predominant carbonates and that from the westward by predominant sulphates. The carbonates eastern water is in the whole of good quality and shows increasing salinity with the length of underground travel. The western sulphated water is good near the northern boundary of the State where the absorption from the Finke and Alberga Rivers keeps it fresh; but the quality deteriorates farther to the south where the gathering ground is more restricted and the conditions for percolation are less favourable.

It is known also that the water of the suspended or marginal basins is of the sulphated type, like that of the main aquifer derived from the western intake.

The following table shows the composition of the water from typical boreholes:—

Analysis	1	2	3	4	5	6	7	8
Ions and Radicles (grains per imperial gallon)	—	—	—	—	—	—	—	—
Chlorine, Cl	17.91	15.74	5.59	25.51	48.50	30.62	67.16	87.27
Sulphuric acid SO_4	Nil	Nil	2.55	Nil	21.96	13.92	34.11	26.13
Carbonic acid, CO_2	37.85	25.35	22.48	45.00	9.45	4.20	16.20	8.19
Sodium, Na Potassium, K	39.66	27.90	21.15	47.64	38.63	20.30	55.44	63.24
Calcium, Ca	Nil	0.79	0.57	1.64	5.14	4.22	8.86	3.50
Magnesium, Mg	0.52	0.43	0.15	0.78	2.42	2.38	3.49	4.20
Silica, SiO_2	0.20	2.50	—	1.00	1.50	1.80	—	1.12
Total saline matter (grains per imperial gallon)	96.14	72.71	52.49	122.17	127.60	77.44	185.26	193.65

Key to analyses :—

1. Lake Letty Bore, southeast of Lake Eyre, water of eastern origin.
2. Kopperamanna bore, east of Lake Eyre, water of eastern origin.
3. Goyder's Lagoon bore, northeast of Lake Eyre, water of eastern origin.
4. Thurlooka bore, east of Lake Frome, water of eastern origin.
5. Oodnadatta bore, water of western origin.
6. Blood's Creek bore, water of western origin.
7. Cordillo bore, water of suspended basin A.
8. Loveday bore, water of marginal basin C.

Variations in the quality of the water obtained from the Jurassic sands of the far south-western portion of the basin are marked by the same features as are possessed by the ground water of western Eyre Peninsula, where there are sediment-filled depressions and valleys in a region of deeply-dissected Pre-Cambrian rocks. The quality is determined by the structural and climatic conditions, although the age of the water-bearing sands is from Tertiary to Recent in the case of Eyre Peninsula.

d) *Diminution of pressure and flow.*

Although there has not been a systematic and regular measurement of the flows from boreholes in the Great Australian Basin isolated observations show that in many cases the present yield is smaller than that recorded when the aquifer was first tapped. In some cases the loss of pressure and flow is attribuable to the collapse or corrosion of the casing. The following table shows the extent of the diminution where it has been ascertained :

Borehole	Date of Drilling	Total Depth (Feet)	Orig. discharge g.p.d.	Most recent measurement Gals. p.d.	Date	Remarks
Clayton	1908	1704	800,000	600,000	1939	
Goyder's Lagoon	1906	4850	600,000	nil	1940	Casing collapsed. water 4ft. below surface.
Kopperamanna	1897	3000	800,000	9,000	1940	Bore partly blocked
Mt. Gason	1903	4420	480,000	450,000	1940	
Marree N° 2	1886	342	84,000	27,000	1891	
Marree N° 3	1908	575	25,000	dribble	1935	
Abdul	1915	341	72,000	21,000	1939	
Oodnadatta	1894	1571	270,000	144,000	1934	
Stevenson	1906	1193	180,000	115,000	1934	
Storm Creek	1898	1550	86,400	74,000	1934	
William Creek. N° 1	1886	272	52,800	nearly stopped	1935	

e) *The temperature of the artesian water*

The temperature of the water discharged from the deepest part of the basin, near the northern boundary of the State is high, and jets of steam are emitted from the boreholes. In these cases temperatures of 208° F (Goyder's Lagoon bore)and 204° F (Mt. Gason bore) have been measured in the bore drains. Doubtless much higher temperatures exist in the aquifer since each of these boreholes is over 4,000 ft. deep. It is noticeable that relatively high temperatures are recorded for the water issuing from shallower boreholes nearer to the margin of the basin, the excess above the normal increment with depth being due, doubtless, to the freedom of movement of the water in the aquifer and the transfer of heat from the deepest portion of the basin by conduction and convection.

There are no thermal springs of normal type, at sites of expiring vulcanism; but the artesian water from the main aquifer does escape at Paralana, on the margin of the basin near Lake Frome, rising along the fault that marks the eastern limit of the Flinders Range, and has a temperature of 144° F.

f) *The development of the basin and the utilization of the water.*

The climatic conditions prevailing in the South Australian portion of the basin are such that there are severe limits upon the number of stock that can be depastured. Consequently there has not been such an insistent demand for many boreholes as has exists in more favoured parts of the continent. Moreover the expense of drilling deeply enough to reach the main aquifer in the far northern part of the state is out of proportion to the stock-carrying capacity of the country. Hence the deepest boreholes have been put by the Government to serve stock routes, and the pastoralists have made great efforts to obtain their requirements from suspended and marginal basin. The following table shows the amount of drilling done in the main and subsidiary basins :—

Boreholes in the Great Australian Basin

Basin	Area sq. mls.	No. of bore-holes	Total Depth of existing bores (ft.)	Remarks
Main Basin, tapping the Jurassic aquifer	120,000	428	187,632	Total flow daily, 28,300,480 galls.
Suspended Basin A	4,035	41	17,163	Almost wholly subartesian
Marginal and Suspended Basin B	3,020	17	8,349	Wholly subartesian.
Marginal and Suspended Basin C	2,498	54	13,319	Wholly subartesian.
Marginal and Suspended Basin D	4,055	53	9,310	Wholly subartesian.
Marginal Basin E	1,484	18	5,736	Artesian and subartesian supplies.

In a few cases the artesian water has been used by the Commonwealth Railways, but the major use is for watering stock, and for this purpose it is almost invariable suitable.

The Murray River Basin

The large part of South Australia over which this basin extends is one of the best favoured portions of the State, with its relatively high and reliable rainfall. The climate, accessibility, and general productivity of much of the region, especially of its southern part, have combined to show that the underground waters of this basin, are at least as important to the State as those of the better known Great Australian Basin in the arid north.

a) *Stratigraphy* and *structure*
The lowest strata are clays (lignitic in part) and sands or gravels with which in places marine beds of Oligocene age are interbedded. These sediments were deposited in a broad gulf that extended formerly beyond the limits of the mapped basin. In the deeper waters a fragmental polyzoal limestone was deposited in Miocene time, and remnants of this formation occur on the highlands to the westward where differential movements in late Tertiary time raised the barrier of the Mt. Lofty Ranges. Between the Murray Bridge-Serviceton railway and the sea there are numerous outcrops of the Pre-Cambrian rocks, mostly granites, which exert a marked control upon the underground water circulation.

b) *Intakes and outlets*
The chief intake areas are situated in counties Lowan and Follett of the State of Victoria, whence the water under pressure moves to the north, northwest and west, as shown by the contours of the hydraulic or piestic surface. There are however, subsidiary intakes on the western margin of the basin where intermittent streams leave the highlands and debouch upon the plains. There are several places on the south coast near Port MacDonnell where the artesian water reaches the surface and is discharged into the sea. Doubtless there is a considerable submarine discharge that is concealed from sight. It has been suggested that there is a discharge also into the River Murray near Bowhill, but confirmation of this suggestion by hydraulic data is not forthcoming at the present time.

c) *The quality of the water*
There is considerable variation in the chemical character of the water in different parts of the basin, the best of it being suitable for all purposes and the worst barely suitable for watering sheep. The best water is that occurring in the far south and the most saline water is that tapped between the River Murray and the Broken Hill railway. In the case of several boreholes drilled in the western part of the basin more than one aquifer exists, and the quality of the water improves with depth, probably for structural reasons, the deeper aquifers being fed by water shed from the highlands and absorbed rapidly by outwash gravels on the upper margin of the plain.

d) *The utilization of the water*
The greater part of the water is used by farmers for watering stock, but supplies have been provided also by the Government for general township use at Mt. Gambier, Naracoorte, Bordertown and Pinnaroo. The paper pulp industry makes use of a large supply derived from a natural vent near Millicent. The number of privately owned boreholes on farmlands is very large.

Since there are comparatively few place at which the water is under sufficient pressure to produce artesian flows where the ground surface is low, there is little waste of the water from artificial openings. A large quantity, however, escapes unused into the Southern Ocean from natural vents near the far southern coast.

The Eucla Basin

a) *Stratigraphy and structure*
The southern coastline of the continent was deeply embayed in mid-Tertiary time and sediments were deposited on the floor of Pre-Cambrian granite or granitic

gneiss. The lowest beds are water-bearing sands and gravels which are overlain by clays (lignitic in part) and polyzoal limestone similar to that of the Murray Basin. The limestone is extremely cavernous.

b) *The character of the water and its utilization*

Most of the water is too saline to be of any use, but in the case of a few boreholes it is barely suitable for watering sheep. In South Australia there are no flowing supplies, but in Western Australia an artesian flow has been obtained from a borehole drilled at a low-level site. Ground or phreatic water is obtainable from shallower depths but in South Australian territory its salinity is high in most cases. There is an improvement towards the Western Australian border, and still better has been found in the western State.

The stock-carrying capacity of the Nullarbor Plain is very low, and the incentive to development has been greatest along the coast where rather better pasturage exists. Exploration for water along the Transcontinental Railway has not been successful in South Australia, so far as quality is concerned.

The Adelaide Plains Basin

a) *The stratigraphy and structure*

This basin is a typical artesian slope situated to the west of the Mt. Lofty Ranges and consisting of a great thickness of sands, clays and limestones which rest upon Pre-Cambrian foundations which have been downfaulted in blocks. The sediments are Mid-Tertiary to Recent in age and aquifers exist at several horizons. The city of Adelaide rests upon a fault block that is notably higher than the foundations farther to the westward ,and artesian water is scanty on the higher block (although. the ground water occurs in places), but fairly abundant where the bedrock is deeper.

b) *The intake and outlet*

Accessions of water to the deeper aquifers are received from the beds of the streams rising in the hihglands and flowing towards Gulf St. Vincent, the principal feeders being the Sturt, Torrens and Little Para Rivers. Percolation to the deeper aquifers probably takes place down the faults on the margins of the fault blocks, notably that known as the Para fault.

There is certainly a submarine discharge or outlet to Gulf St. Vincent but no other natural outlet. Although artesian conditions prevail in the lower country bordering on the coast the flow from artificial openings is small unless assisted by pumps.

c) *The utilization of the water*

The stored water has proved to be of great value to the metropolitan area in times of drought, when reservoirs have been seriously depleted; and large quantities have been pumped into the mains supplying the city of Adelaide and its suburbs. At such times it has been the practice to draw only from those aquifers which furnish water of good quality, carrying between 47 and 68 grains per imperial gallon of total dissolved solids. Much of the artesian water has been used for industrial purposes and in horticulture.

The Pirie-Torrens Basin

a) *Structure and stratigraphy*

This basin resembles in a general way the Adelaide Plains Basin, being an artesian slope occupying a piedmont plain to the westward of the Flinders Range. Upper Tertiary to Recent sediments rest upon Pre-Cambrain down-faulted blocks, and these are several aquifers in the sub-horizontal sediments, fed in the same way as the aquifers of the more southerly basin.

b) *The utilization of the water*

There are many boreholes providing stock water, and a few which furnish supplies for horticultural use. The latter are situated opposite the mouths of intermittent streams rising in the highlands.

Minor Artesian Basins

The geological features exhibited by the smaller and less important basins are normal and call for no special comment. The variations in the quality of the water contained in the different basins are due to differences in the nature of the intakes. Where these are limited and the percolating water has suffered excessive evaporation, soil salt is taken down into the aquifers. But, where there is rapid absorption and relatively little evaporation after the water leaves the gathering grounds on the highlands, the quality is appreciably better.

SUPPLIES OF GROUND (OR PHREATIC) WATER

The most important ground water supplies, from the viewpoint of the community generally rather than from that of the individual, are those of consideralbe size and broad lateral extent such that they can supply the demands of many. The outstanding examples of such supplies are known in different parts of the State.

a) The southeastern groundwater

Overlying the Oligocene sands and gravels of the south-eastern region are porous limestones and marls, with peaty soil cover in low-lying places. These porous limestones carry groundwater that is known to be moving steadily in a westerly direction from Victoria towards the Southern Ocean. It has not been developed extensively in the past, but with the progress of closer settlement in this region considerable use will be made of it in the future. Recently a considerable number of boreholes has been drilled to equip now settlers with their own supplies. The boreholes thus drilled number 121, and they range in depth from 14 to 235 feet, the average depth being 71 feet. The average tested pumping yield from these boreholes is 617 gallons per hour.

The ground-water of this area merges with the underlying artesian water to the south of a major structural break known as the Tartwaup Fault. The mergence of the waters is influenced also by the volcanic vents which have traversed all the sediments. The lakes in the volcanic craters at Mt. Gambier are regarded as exposed portion of the water table.

b) The Flinders or Streaky Bay ground water

A shallow basin, about 200 square miles in extent, carrying sandstone capped with travertinous limestone, on Pre-Cambrian foundations, has been found to contain useful water resting on much more saline water. Special measures were adopted to draw upon the top water only, with its content of 40 to 50 grains of total dissolved solids per imperial gallon, and to prevent contamination by the underlying water that carries 3 oz. or more of dissolved solids per imperial gallon. These measures have been successful and it has been possible to pump over 15 million gallons per annum without depleting the storage.

c) The Uley-Wanilla ground water

This basin is relatively small, comprising only 20 square miles, and is situated about 15 miles to the north-west of Port Lincoln. The basin is an eroded hollow in the Pre-Cambrian rocks and is filled with wind-blown shelly sand capped by travertinous limestone. The porosity of the sand is 28 per cent, and its maximum depth is 80 feet. The principal part of the basin is estimated to store over 7,000 million gallons of water which contains between 20 and 30 grains of dissolved solids per imperial gallon. The average rainfall at this place is 23 inches per annum and if 25 per cent of that rainfall reaches the water table the annual replenishment is expected to be over 900 million gallons.

d) Other groundwater basins

A broad valley filled with calcareous sand exists in county Robinson, on western Eyre Peninsula and carries a considerable amount of water westwards towards Elliston. The quality of the water is variable and much remains to be done before the value of this basin can be assessed.

There is a large area between the Alberga River and Hamilton Creek in the far northern part of the State that carries useful stock water near creeks that bring occasional accessions of flood water. Little use is yet made of this basin.

e) Eyre Peninsula groundwater

In central and western Eyre Peninsula, beyond the limits of the basins to which reference has been made above a maturely-dissected granitic or gneissic terrain has had the hollows filled with sandy sediments since early Tertiary time. Percolation is marked, and no surface streams exist. The deep groundwater at the bottom of the sediments is saline save for a « cream » of better water occurring upon the salt. Useful supplies have been obtained from wells and boreholes that have reached the foundation rocks above the level of the deeper salt water.

f) Groundwater in sand dunes

Some of the earliest supplies obtained and used by settlers coming to South Australia were derived from shallow wells sunk in the sands close to the shore. A few of these wells are in use today.

Similar supplies have been drawn from inland dunes, the most notable being that located at Ooldea on the Transcontinental Railway and on the eastern margin of the Eucla Basin.

The older, Pleistocene, dunes near the coast exhibit features similar to those of the present day, differing chiefly in their consolidation by solution and redeposition of the calcareous matter entering into their composition. Useful water has been obtained from these dunes to the north-west of Fowler's Bay, the water table being controlled by sea level. There is a « cream » of better water above more salinewater in these wells also.

g) Groundwater in dense rocks

Many supplies, suitable only for individual farms, are derived from wells and boreholes sunk or drilled in rocks traversed by joints fractures or other openings that serve to admit and store water. An exceptional case is that of the Bird-in-Hand mine at Woodside in the Mt. Lofty Ranges where it was necessary to pump out a million gallons a day to enable mining operations to be carried on. The water was stored in a fault zone partly filled with vein quartz, and this yield was maintained for a period of 12 months.

It has been found in South Australia, as elsewhere, that limestones are more favorable than other dense rocks as sources of water, and it has been proved also that anticlinal folds are likely to yield large supplies. One example is that of the underground supply used for the Leigh Creek coalfield, which is connected by a 25-mile pipeline with a borehole near the Sliding Rock mine, drilled in Cambrian limestone.

S. B. DICKINSON
Government geologist

Free translation from the Spanisch of a summary of grond water in Argentina
submitted by **Senor Juan Hugo CAESAR**, March 10, 1948

7) GROUND WATER IN THE ARGENTINE REPUBLIC

Province	*System*	*Ground-water conditions*
Buenos Aires :	Quaternary	Fresh and brackish water aquifers of moderate yield. Quality of water varies with individual aquifers.
	Tertiary	*Pliocene* : In the Puelchenese formation aquifers yield adequate supplies of water generally of fair quality but infrequently brackish. Aquifers generally unconfined but locally may be confined. Araucano formation aquifers yield moderate supplies generally brackish water. *Miocene* : Green clay beds : Aquifers yield adequate supplies of salty water. Aquifers confined. Red clay beds : Aquifers yield adequate supplies of water generally salty but sometimes potable. Aquifers confined.
Catamarca :	Quaternary	In piedmont and bolson deposits distributed by faults are aquifers that yield small to moderate supplies of fresh and brackish water. Aquifers generally unconfined but locally may be confined.
	Tertiary	In the Calchaqueno strata aquifers yield moderate supplies of salty water. Aquifers unconfined. In the Araucano formation individual aquifers may yield moderate supplies of fresh or salty water. Aquifers unconfined. *Miocene* : In the Llanos strata aquifers yield small to moderate supplies of salty water. Aquifers unconfined.
La Rioja :		Same as in Catamarca province.
Cordoba :	Quaternary	In alluvial fan deposits aquifers yield small to moderate supplies of water of good quality. Aquifers unconfined.
	Tertiary	*Pliocene* : Aquifers yield small supplies of fair brackish water. Aquifers unconfined. *Miocene* : Aquifers yield moderate supplies of water of poor quality. Aquifers locally confined.
Corrientes :	Quaternary	In deposits of limited areal extent and thickness aquifers yield small supplies of water of good quality.
	Tertiary	*Pliocene* : Aquifers yield small supplies of water of good quality. Aquifers unconfined.
	Triassic	Melaphyre yields moderate supplies of water of good quality. Aquifers unconfined.
Entre Rios :		Same as in Buenos Aires province without the red clay beds.
	Triassic	Melaphyre yields small supplies of water of good quality. Aquifers unconfined.
Jujuy :	Quaternary	In alluvial fan deposites aquifers yield small to large supplies of fresh or brackish water. *Cretaceous to Pliocene* : Aquifers yield moderate supplies of generally brackish water. Aquifers unconfined.
Mendoza :	Quaternary	In alluvial fan deposits aquifers yield adequate to moderate supplies of water that is generally of good quality. Aquifers locally confined.

	Tertiary	*Pliocene* : In Calchaqueno strata aquifers yield small to moderate supplies of water that is generally brackish but occasionally fresh. Aquifers unconfined.
Salta :	Quaternary	In alluvial fan deposits aquifers yield small to abundant supplies of water of good quality. Aquifers unconfined.
	Tertiary	*Pliocene* : Aquifers yield moderate to adequate supplies of fresh or brackish water. Aquifers generally unconfined but locally confined.
	Triassic to Cretaceous	Aquifers yield moderate supplies of brackish water. Aquifers unconfined.
San Juan :		Same as in Mendoza province.
San Luis :	Quaternary	In alluvial fan deposits aquifers yield small to moderate supplies of fresh to brackish water. Aquifers generally unconfined.
	Tertiary	*Pliocene* : Aquifers yield moderate supplies of water that is generally brackish but occasionally fresh. Aquifers generally unconfined., but locally confined. *Miocene* : Aquifers yield adequate supplies of water generally of good quality. Aquifers locally confined.
Santa Fe :		Same as in Buenos Aires province.
Santiago del Estero :		In this province aquifers in formations from the Miocene to Quaternary, inclusive, generally yield brackish water. Locally aquifers in the Quaternary may yield fresh water. Aquifers yield small to moderate supplies and are generally unconfined.
Tucuman :	Quaternary	In alluvial fan deposits aquifers yield moderate supplies of water of good quality. Aquifers locally confined.
	Tertiary	*Pliocene* : Aquifers yield moderate supplies of water of good quality. Aquifers confined.
Formosa :		Aquifers in the Quaternary and Pliocene yield small to moderate supplies of brackish water. Aquifers unconfined.
Chaco :		Same as in Formosa province also including the Miocene.
La Pampa :		Aquifers in Miocene to Quaternary formations yield small supplies of generally brackish water. Aquifers unconfined.
Misiones :		Same as in Corrientes province. Aquifers in melaphyre.
Rio Negro :		Generally aquifers in formations from the Cretaceous to Quaternary yield small to moderate supplies of water of good quality. Aquifers confined.
Chubut :	Quaternary	Aquifers practically absent.
	Tertiary	« *Patagonic* » : Aquifers yield moderate supplies of water of good quality. Aquifers unconfined. *Lower Miocene* : Aquifers yield moderate supplies of brackish water. Aquifers unconfined.
Santa Cruz :		Same as Chubut province.
Neuquen :		Same as Chubut province.
Federal Capital		Same as Buenos Aires province.

COMMUNICATIONS
DIVERSES

1) TAKING OF SOIL PROFILE SAMPLES

by

MATTI WÄRE

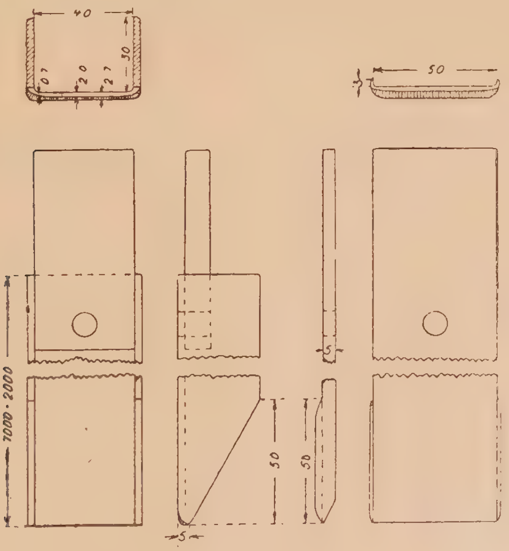
At the Experimental Station for Agricultural Hydrology at Maasoja (Wihti, Finland) it was considered that the methods developed in Germany for taking soil samples were not satisfactory, as comparisons of results obtained with different tools showed very great differences.

After many trials it was found that the equipment described below made it possible to take good profiles samples (up to 160 cm in lenght), both from peat soil and from dence clay soil, and also from frozen soil.

The tool used is made of ordinary building industry profile steel. The cross-section of the soil profile sample is 3 × 4 cm, its length is 70 cm when using a tool 100 cm long, and 160—170 cm when using a tool 200 cm. long. One side of the tool is uncovered. — It is driven down with a small hammer. When the necessary depth has been reached, the open side is covered by driving down a cutting steel chisel, the projections of which prevent it from slipping sideways. The soil profile sample is raised by forcing up the tool and chisel at the same time The sample has been cut in lengths of 5 cm, and thus it has been possible to determine the soil separately for each of these pieces of 60 cm^3.

Comparisons have shown that the water content of the soil has been determined with an exactness equalling that obtained by using Kopécky cylinders of 180 cm^3. Our soil profile samples have not disturbed agricultural work, as they cause only a small opening (3 × 4 cm) on the soil surface.

It has been possible to use the sample also for determining the water capacity of the soil. For this the open side is covered with a brass-wire net, the long sample in the tool being laid horizontally at the disired distance from the water surface, with which a blotting paper contact has been established. It has been found that the accuracy equals that obtained by using Kopécky cylinders.



2) ABOUT SOIL HUMIDITY VARIATIONS

by

MATTI WÄRE

At the Experimental Station for Agricultural Hydrology in Maasoja (Withi, Finland) a sum total of 1019 soil profiles were taken between 1939 and 1944 with the aid of the tool described in another paper (Taking of Soil Profile Samples; by the Author). These soil profiles were cut in 11,667 different pieces (3 × 4 × 5 cm) and thus a considerable material was got together for determining the water contents of the soil at different depths, in different kinds of soil (clay or peat, whether drained by open ditches or by covered drains), and during different seasons.

On a field drained by open ditches (depth of the ditches being 60 cm) the humidity variations were in general confined to the upper stratum of 70 cm thickness.

In dense fine clay soil the highest and lowest humidity values were to be found in the surface layer 0—20 cm. In very slimy peat soil the humidity values of the top strata never reached those of the lower layers, although the variations were very great near the surface.

There was in early spring a soil humidity maximum amounting to 52% (profile value for 0—60 cm, average for 1941-44) in a clay field drained by open ditches, and to 46 % in a clay field drained by a covered drainage system (pipes 110 cm below the surface). Corresponding values for peat soil were 62 % (open ditches) and 53 % (pipe drained).

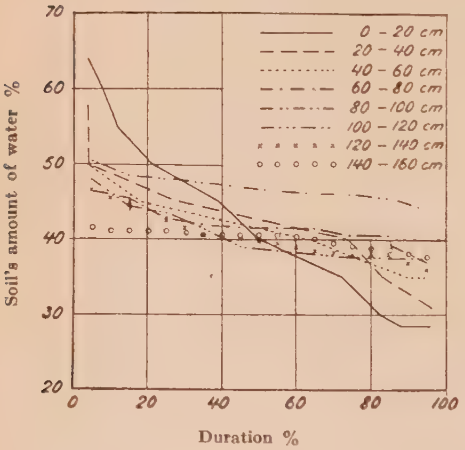
Thus the humidity has been reduced 4-9 percentage units during the spring maximum and the summer minimum, if pipe drains have been used instead of open ditches.

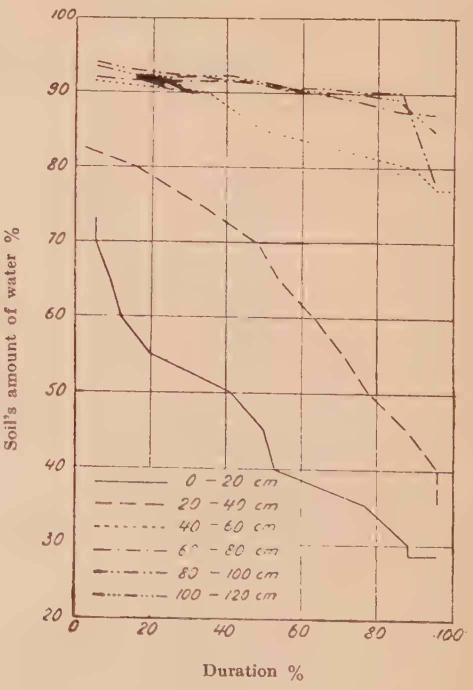
By consulting duration curves, it was found that this same difference prevailed continuously throughout the year, not only during the extreme periods.

The water content difference between peat soil and clay soil was 22 percentage units during the spring maximum, and 25—30 percentage units during the summer minimum.

In a surface layer of 20 cm thickness the humidity difference between peat and clay soil was only 5—10 percentage units in winter, and still less in summer. On the other hand, in the layer 40—60 cm from the surface the humidity difference amounted to 40—50 percentage units.

Cf. MATTI WÄRE, Maan vesisuhteista ja viljelyskasvien sadoista Maasojan vesitaloudellisella koekentällä vuosina 1939—1944, *Referat : Uber die Wasserhältnisse des Bodens und die Erträge von Kulturpflanzen auf dem wasserwirtschaftlichen Versuchsfeld Maasoja in den Jahren 1939—1944*, Maa- ja vesiteknillisiä tutkimuksia N:o 5, Helsinki 1947.





3) THE EFFECT OF THE DRAINING PROCEDURE ON THE HUMIDITY OF THE SOIL, THE FROST AND THE GROUND TEMPERATURE

by

TANELI JUUSELA

In connection with tests concerning the effect of the draining method (pipes or ditches) on the humidity, the freezing and the soil temperature, comparing observations were made during the period 1.6.1940—25.8.1942 on the Backas Estates of Helsinki parish. The soil was hard clay. The tests concerned four different conditions and the observation points were situated in the centres of tilled plots as well as on their borders, at a distance of 1 meter from open ditches or pipes.

Underground water level and humidity. It was established that the underground water level in spring lay deeper in the pipe-drained plots than in the plots with open ditches. In early summer — June — the underground water level was in each of the three years sinking quicker in the field with open ditches than on the pipe-drained field. During the exceedingly dry and hot summers of 1940 and 1941 this level remained higher in July as well as in August on the pipe-drained field than on the field with ditches. This fact together with the humidity analyses of soil samples proves that the pipe-drained soil retained its humidity better than the soil drained by ditches. This is also rather natural, as the continuous granular surface of the pipe-drained soil is apt to prevent the evaporation, which on the other hand will be furthered by open ditches. The estimates based on observations of the underground water level and of the soil humidity plainly show that for instance in June 1940 the evaporation on the field with open ditches, was 40 percent higher than on the pipe-drained field where the soil humidity already in the middle of June (in 1940 as well as in 1941) was greater than on the field with ditches.

On the other hand, during its maximum epoch in the autumn, and for a time succeeding this epoch, the ground water level lay deeper in the pipe-drained field than in the field drained by ditches, thus showing a return to spring conditions.

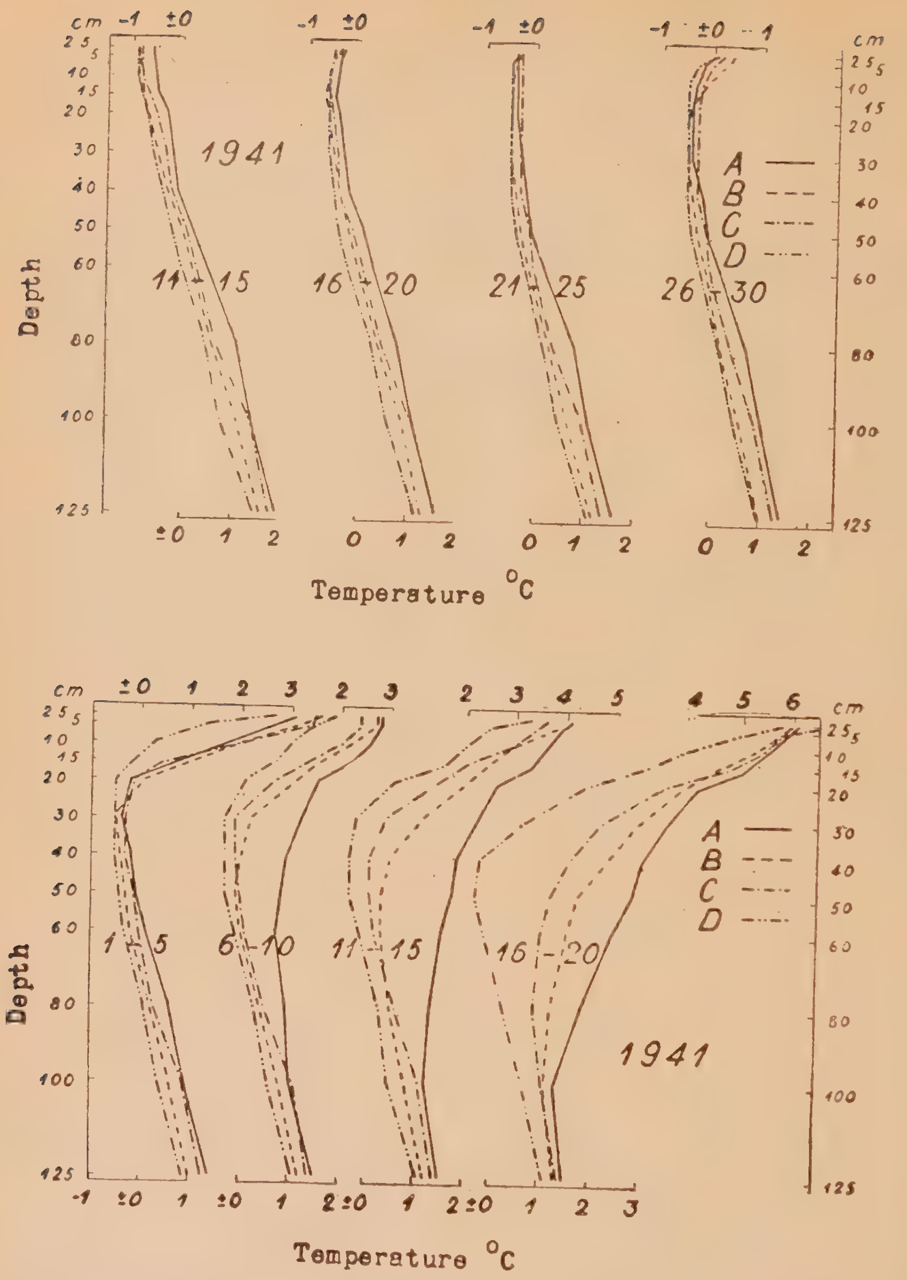
The ground frost. The investigations proved that the frost would go deeper into the soil with open ditches than into the pipe-drained one. Thus the depth of the ground frost was in the late winter of 1941 15 percent greater in the field drained by ditches than in the field drained by pipes; corresponding figure for the late winter of 1942 (when the soil was frozen 20-30 cm deeper than in the preceding winter) was 20 percent.

The water percentage of the frost strata was greater on the field with open ditches than on the pipe-drained field, or, generally speaking : the humidity percentage was the greater, the higher the underground water level had been during the time of freezing,

The thicker and more moisture containing frozen strata on field with open ditches meant a delay of the melting in the spring. The observations made during the melting period in the spring of 1941 showed that the frozen strata on the pipe-drained field melted 7 to 10 days earlier than on the other field.

The ground temperature. In order to pursue the fluctuations of the ground temperature, copper-constantan thermo-elements were employed by using the so-called compensation method. The ground temperature was measured for the following depths : 2 ½, 5, 10, 15, 20, 30, 40, 50, 60, 80, 100, and 125 cm. The air temperature in the vicinity of the observation points and in the heights 10, 50 and 200 cm above the ground surface was also determined by using thermo-elements.

The temperature observations showed that practically the same temperature prevailed on the different observation posts in the autumn when the humidity fluctuations were very small. The temperature differences remained negligent in the early winter as long as the ground was not frozen. When freezing however set in the temperature differences grew greater and during the cold period the ground temperature on the different sites of observation was the lower the more moisture there was in the frozen layer and the higher the ground water level had been in the autumn. Thus the observation points on the pipe-drained fields showed a higher ground temperature than on the fields with open ditches.



The ground temperature during the melting process in April (upper figure) and in May (lower figure) in the year 1941.

Observation point A = at a distance of 1 meter from pipe-drain
» » B = halfway between pipes
» » C = at a distance of 1 meter from open-ditch
» » D = halfway between ditches.

In spring when the ground temperature rose approaching the zero the temperature fluctuations in the frozen layer as well as below it were reduced, being at the time of melting almost as insignificant as during the autumn months. By the melting process the ground temperature in the frozen layer rose all the slower, the bigger the ice volume was in the frozen stratum, and consequently quicker on the pipe-drained fields than on the fields with open ditches. Thus the ground on the observation sites of the pipe-drained field in a depth of 20-60 cm was in 1941 around May 15th, when the temperature differences reached their maximum values, 2.4° C warmer than the ground on the ditchdrained field.

The melting time high values of the temperature differences in pipe-drained and ditch-drained soil decreased more rapidly in the rainy summer 1942 than in the hot and dry summers of 1940 and 1941. At a depth of 20—40 cm these differences were in June and July of 1940 and 1941 about 1.5° C, while in 1942 the difference was about the half of the above value in June and had almost completely vanished in July.

With reference to the effect of the humidity of the soil upon its temperature, the experiments show that the ground temperature will follow the fluctuations of the air temperature and this will happen more closely if the humidity values are high. This result could be explained in the following way : With an increase of humidity the heat conductibility of the clay ground will increase. The effect of this higher conductibility may not only compensate but also overcome the delaying effect of the greater water contents on the temperature variations.

Cf. TANELI JUUSELA. Untersuchungen über den Einfluss des Entwässerungsverfahrens auf den Wassergehalt des Bodens, den Bodenfrost und die Bodentemperatur, Acia Agralia Fennica 59, Helsinki 1745.

4) ISO-RESISTIVITY MAPPING FOR THE INVESTIGATION OF UNDERGROUND WATER SUPPLIES

by

G. L. PAVER,
M. B. E., M. A. Ph. D., A. I. M. M., F. G. S. Member, American Society of Exploration Geophysicists

Introduction
Geophysical Technique
Water Table Conditions in Mio–Pliocene Sediments
Water Table Conditions in Eocene Limestones
Zone of Decomposition in Pre-Cambrian Quartzites
Zone of Decomposition in Pre-Cambrian Lavas
Fissured Zone in Pre-Cambrian Granitic Basement Complex
Fracture Zone in Association with Basic Igneous Intrusion in Granite

Introduction

The proper scientific investigation of the problems of underground water supply should embrace the correlation and reconcilement of all available and observable hydrological, geological and geophysical data. The geophysical work should rest on a sound geological and hydrological foundation and be regarded as an extension of the collection of attainable data and not looked upon as a separate investigation. No interpretation of observed results, however feasible mathematically or physically, are acceptable if they contradict the established facts of hydrology and geology. The initial application and final interpretation of geophysical data demands a thorough knowledge of geology and an understanding of the hydrological conditions controlling the occurrence of underground water in the various rock types.

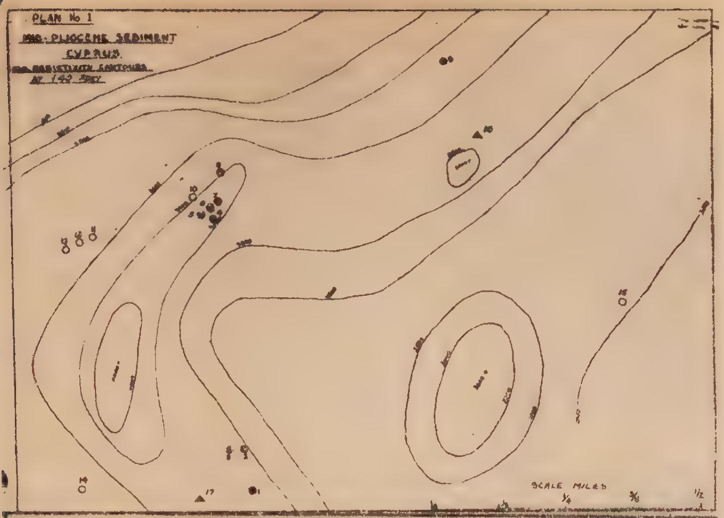
Geophysical Technique

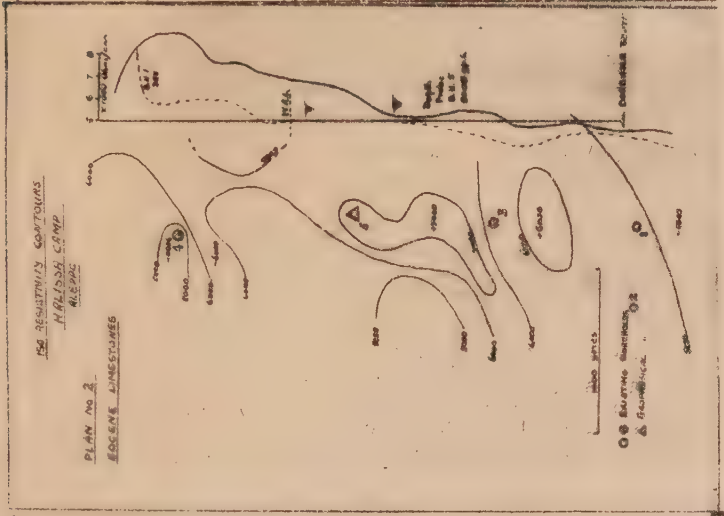
The geophysical investigation of underground water makes full demands on geological knowledge and field ability and also requires a speed of operation capable of producing successful results in conjunction with a time factor such as to render the use of geophysics an economic proposition. The comparatively laborious gravimetric and seismic methods are of importance in the investigation of regional or deep-seated water supply problems, but the established method applied to the usual problems of water supply is that of electrical resistivity.

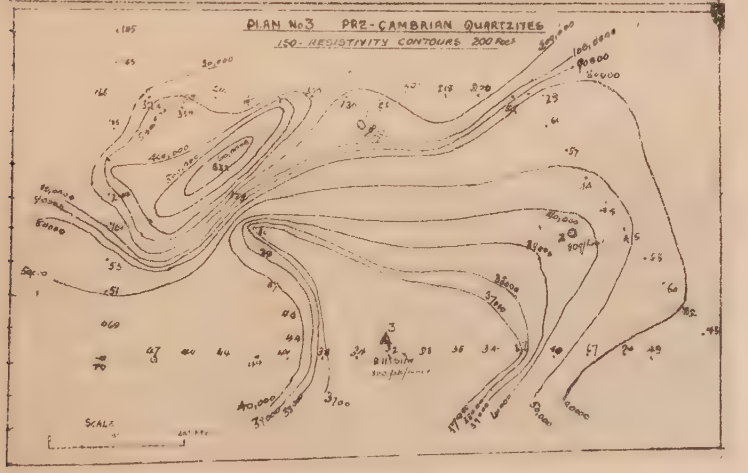
Electrical resistivity surveys are not only comparatively economical and speedy in operation, but the water content of the rock is one of the factors directly affecting observed values and direct evidence of underground water conditions is therefore partially attained. The electrical depth probe curve constitutes the basis of the plotted resistivity results, while reconnaissance observations may be plotted as constant separation profiles and final results plotted as iso-resistivity diagrams at varying depths.

Iso-resistivity plans are prepared by interpolation from a number of depth probe measurements or by the observation of a grid of constant depth measurements at a selected depth dependant upon known or reasonably assumed hydrological and geological factors. The iso-resistivity map has certain advantages in interpretation and, where surface factors are not unfavourable, in speed of operation; trends of resistivity depth variations are quickly discernable which are often only with difficulty interpreted from individual depth probe curves.

The following examples of iso-resistivity mapping in limited areas of proved difficult underground water conditions are selected both as fairly representative of the occurrence of underground water in various rock formations as well as to demonstrate the comparative advantages of the technique of iso-resistivity mapping.


PLAN No 1
MIO-PLIOCENE SEDIMENT
CYPRUS.
RESISTIVITY CONTOURS
AT 140 FEET
SCALE MILES
1/4
1/2


150 RESISTIVITY CONTOURS
HALISSH CAMP
ALEPPO
PLAN No 2
EOCENE LIMESTONES
EXISTING BOREHOLES
GEOPHYSICAL


PLAN No3 PRE-CAMBRIAN QUARTZITES
150- RESISTIVITY CONTOURS 200 Feet
SCALE

Water Table conditions in Mio–Pliocene Sediments

The area depicted in this iso-resistivity plan lies immediately to the North of the railway station at Nicosia, Cyprus. Geologically the area is one of Mio–Pliocene sands, gravels and alluvials; hydrologically existing wells and bores had already established the general water table rest level to be at 85 feet below ground level. Seventeen boreholes are in the area under consideration; of these fifteen existed prior to the geophysical work, nine of which were successful in producing strong supplies and six of which were failures. The producing boreholes having penetrated sands, gravels and conglomerates below water table level yielded 1000-3000 g.p.h., while the failures invariably resulted from clays being encountered in boring. The geophysical problem therefore was to delineate maximum resistivity zones representing the water yielding sands and gravels below water-table level.

Plan nº 1 shows the iso-resistivity map at an apparent depth of 140 feet as interpolated from the results of some twenty four depth probe measurements observed-roughly in grid form at quarter mile intervals. There is of course room for localised variations, but the general trend of the depth resistivity variations is quite clear and indicates a comparatively higher resistivity zone, representing coarser grained potentially water-yielding sediments running in an approximately SE-NWdirection.

Of the nine originally successful boreholes in this area, of approximately 1 ½ miles by 1 mile, eight were found to lie in the zone of observed resistivity greater than 3000 ohm /cms at the apparent depth of 140 feet while the other, nº 9, lies close to it; only one unsuccessful bore is situated within this resistivity zone and the depth probe at this site tends to indicate that it is located in a limited area anomalous to the general trend. The other four unsuccessful boreholes all correspond to the 140 foot depth resistivity range of below 3000 ohm /cms.

Two further borehole sites, Nos. 16 and 17 were drilled as a result of the geophysical work and yielded 1800 and 1200 g.p.h. respectively.

Water Table Conditions in Eocene Limestones

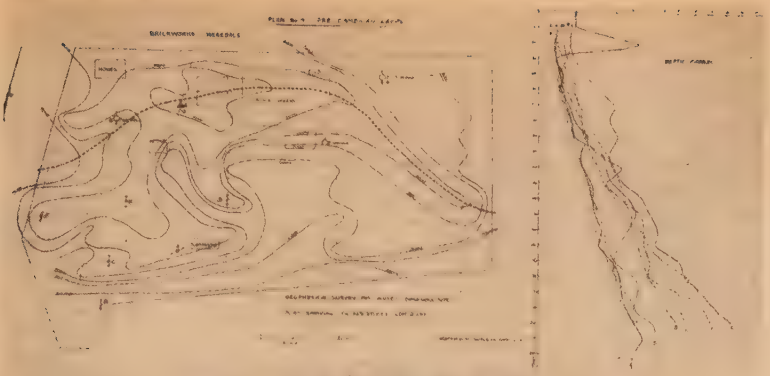
The limestone formation under consideration is in the Aleppo area, Syria. The average rainfall is 17″ per year; the geological conditions are those of Eocene limestones varying in testure, both laterally and vertically, from comparatively impermeable marly limestones to the more water productive silicified limestones; hydrologically, it was establisjed from observations in wells and existing boreholes that the water table level was remarkably consistent at 418 Metres above sea level.

With these conditions obtaining, the geophysical problem was to locate maximum resistivity zones at water table level which could be interpreted to indicate the presence of the more water yielding silicified type of limestone. Iso-resistivity plan No. 2 shows the results of the geophysical survey plotted at an apparent depth of equivalent to water table level, the effects of a somewhat undulating topography being taken into account. It can clearly be seen that the four original boreholes two of which were dry and two of which produced small yields only, are in the 5000 ohm /cms zone, the geophysically selected site, which was successful in producing 5000 g.p.h. from a similar completed drilling depth, is in a zone of resistivities above 7,000 ohm /cms at water table level.

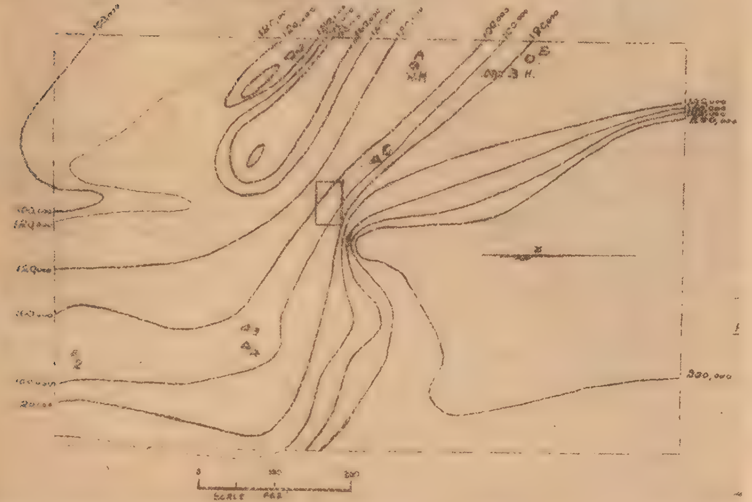
Decomposed Zone in Pre-Cambrian Quartzites

The iso-resistivity plan No. 3, plotted from a grid of constant apparent depth measurements at 200 feet covers an area of 10 acres, geologically consisting of comparatively coarse grained, but hard and compact Elsburg Quatrzites of the Upper Witwatersrand System, S. Africa. No igneous intrusion or faulting and associated fissuring favourable to underground water conservation was present and water, to exist, would have to be in conjunction with open bedding planes or optimum area and depth of decomposition. That the former was not the case was proved by the completion of two dry boreholes prior to the commencement of geophysical work.

The iso-resistivity plan clearly delineates a favourable low resistivity zone of below 37,000 ohm /cms at 200 feet depth and shows the previous bores particularly bore No. 1, not to have been well situated. The borehole drilled at the geophysical site No. 3 produced the adequate yield, for such a formation, of 750 g.p.h. at a completed depth of 225 feet.



ISO-RESISTIVITY PLAN 5 : GRANITES
AT 100 FEET DEPTH



PLAN No 6 NORITES.

Iso resistivity Contours

A at existing borehole
B & C at proposed borehole sites

Fissured Zone in Pre-Cambrian Basement Complex

Fractured and fissured zones in igneous rock represent potentially water-producing conditions in otherwise massive, crystalline formation typical of igneous geology particularly the holocrystalline plutonic rock types. Such conditions may be rapidly ascertained and outlined by resistivity mapping and are represented by low resistivity values. Plan No. 5 represents the results of iso-resistivity mapping covering an area of some 10 acres. Two boreholes, one with a small supply, which dried up, and one drilled to 105 feet in solid granite and being quite dry, were in existence at the commencement of geophysical survey work.

The iso-resistivity plan plotted from a grid of measurements at 150 feet in depth and a station interval of 50 feet shows the area to be one of high depth resistivity values, correlatable with massive granitic conditions unfavourable to the presence of underground water. A well marked narrow zone of comparatively low resistivity values below 100,000 ohm /cms is clearly indicated to run through the area in question. The existing boreholes are situated on either side of this zone interpreted geologicaly to represent fissuring associated with a near-by major quartz crush zone. The first geophysically selected site drilled within this fracutre as localied by the iso-resistivity plan and confirmed as a favourable site by depth probe measurements yielded 850 g.p.h. at a completed depth of 130 feet which represents an above average yield for this rock type in the area concerned.

Decomposed Basin in Norites

In common with other plutonic igneous rocks areas of maximum decomposition to maximum depth represent optimum locations for drilling for water. The more basic norites show a lower resistivity range than their granitic equivalents but iso-resistivity mapping is again capable of quickly indicating resistivity variation trends and isolating decomposed basins in the otherwise non-porous crystalline rock as indicated by minimum resistivity zones.

Plan No. 6 represents the iso-resistivity plan of an area within the Norites of the Bushveld Igneous Complex. The original borehole yielded a medium supply of 350 g.p.h. and is situated in the 17,000 ohm /cms zone ; the geophysically selected borehole C in the low minimum resistivity area of 8,000 ohm /cms yielded 750 g.p.h. the borehole at B however was surprisingly unsuccessful whereas boring at D gave similarly 700 gallons per hour, these latter both being situated in the 10-11,000 ohm /cms zone.

Conclusion

In areas where detailed investigations into underground geo-hydrological conditions and water-bearing potential are required, the technique of iso-resisitvity mapping has certain practical and interpretative advantages. Careful resistivity measurements interpreted in the light of established geological and hydrological facts may at times lead to highly satisfactory borehole results in areas otherwise appearing to be comparatively unproductive of underground water supplies.

TABLE OF BOREHOLES

Plan	Borehole	Formation	Depth in feet	Yield in g.p.h.
1	1 - 9	Mio–Pliocene Sediments	150 - 200	1000-3000
	10 - 15	do.	» »	Poor
	16	do.	182	1800
	17	do.	150	1200
2	1	Eocene Limestone	150	nil
	2	» »		
	3	» »	210	350
	4	» »	200	240
	5	» »	237	5000
3	1	Pre-Cambrian Quartzites		nil
	2	do.		80
	3	do.	225	750
4	1	Pre-Cambrian Lavas	200	60
	Windmill	do.	200	600
	D	do.	350	Nil
	E	do.	350	Nil
	A	do.	165	850
5	A	Norites	145	300
	B	»	200	Nil
	C	»	165	750
	D		165	700
6	A	Granitic Basement Complex	145	Dry
	B		105	Nil
	C		130	850

5) GRUNDWASSER-HAUSHALT

UNTERSUCHUNGEN IM GRUNDWASSERSTROM DES REUSSTALES IN QUANTITATIVER UND QUALITATIVER HINSICHT.

von

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Die chemischen Untersuchungen wurden in meinem Auftrage durch das Laboratorium der Beratungsstelle für Trinkwasserversorgung und Abwasserreinigung der Eidg. Techn. Hochschule in Zürich ausgeführt.

A. Allgemeines :

Etwa 10 km stromabwärts ihres Ausflusses aus dem Vierwaldstättersee beschreibt die Reuss einen Bogen nach Nordwesten, welcher durch den 3,1 km langen Wasserwerkskanal der Papierfabrik Perlen abgeschnitten wird. Der Bogen des Reusstales an jener Stelle liegt noch im Gebiet des eiszeitlichen Sees, welcher sich von Bremgarten bis oberhalb Perlen erstreckt haben muss. Auf den schlammigen Lehmboden dieses Sees sind dann in einer späteren Phase die groben und durchlässigen Kiese verfrachtet worden, welche von der Endmoräne des Reussgletschers bei Luzern stammen. In diesen, den eigentlichen Talboden in einer Stärke von 8 - 13 m füllenden Kiesträger, findet in der Gegend von Emmen eine starke Infiltration durch die Reuss statt, so dass sich von dort an ein starker Grundwasserstrom bildet, der bei Perlen schon über die ganze mehr als 1km betragende Talbreite festgestellt werden kann. Die lokale Einengung des Talprofils nordwestlich Root im Zusammenhang mit einem leichten Ansteigen der undurchlässigen. lehmigen Sohle des Grundwasserträgers bewirkt ein Austreten eines Teiles des Grundwasserstromes an die Oberfläche, so dass, sich in der Gegend von Perlen mehrere Grundwasseraufstösse bilden, welche in verschiedenen Gerinnen oberflächlich abfliessen und weiter stromabwärts entweder wieder in den Kiesträger versickern, oder sich direkt in die Reuss ergiessen.

Die allgemeine Situation des Grundwassergebietes von Perlen ist in Abbildung 1 ersichtlich. In diesem Plan sind auch die Gebäude der Papierfabrik Perlen eingezeichnet. welche seit ihrem Bestehen das notwendige Trink- und Betriebswasser dem Grundwasserstrom des Reusstales entnimmt.

Die zwei ersten, als Kesselbrunnen gebauten Fassungen A und B, welche ganz am rechtsseitigen Talhange liegen, zeigten sich der ständig zunehmenden Betriebswassermenge der Fabrik bald nicht mehr gewachsen und wurden in den Jahren 1912 und 1915 durch die beiden Rohrfilterbrunnen I und II ergänzt, welche schon weit näher dem eigentlichen, ergiebigen Grundwasserprofil des Talbodens liegen. Durch die ständig zunehmende Vergrösserung der Fabrik wurden aber auch diese zwei Rohrbrunnen bald überbeansprucht. Im Winter 1927-1928 wurde erstmals ein starker Eisengehalt im geförderten Betriebswasser bemerkt, welcher mit den verschiedenen Wasserständen schwankend, 0,3 bis 0,8 mg/l betrug, wogegen z.B. Klut in seiner « Untersuchung des Wassers an Ort und Stelle », 5. Auflage, Seite 66, angibt, dass für die Herstellung sehr feiner weisser Papiere nur ein nahezu eisenfreies Wasser mit höchstens 0,05 mg/l, brauchbar ist. Im Jahre 1928 wurde daher der ganze zwischen Fabrik und dem Fendlibach-Giessen liegende Teil des Grundwasserträgers durch eine Gruppenfassung von 4 modernen Rohrbrunnen A_1 bis A_4 erschlossen, denen dann im Jahre 1934 noch ein weiterer Rohrbrunnen A_5 nachgeschaltet werden musste.

Durch die oben erwähnten periodisch aufgetretenen Verschmutzungserscheinungen gewarnt, liess von nun an die Fabrikleitung in regelmässigen Intervallen die Oxydierbarkeit des gepumpten Wassers durch Feststellung des Kaliumpermanganat-Verbrauches ($KMnO_4$) ermittlen, welche einen allgemeinen Indikator für den Gehalt des Wassers an reduzierenden Stoffen darstellt. Die Schwankungen in der Oxydierbarkeit des gepumpten Grundwassers sind in Abbildung 2 vom April 1935 bis und mit April 1937 aufgezeichnet. Währenddem die Verschmutzungen im Ver-

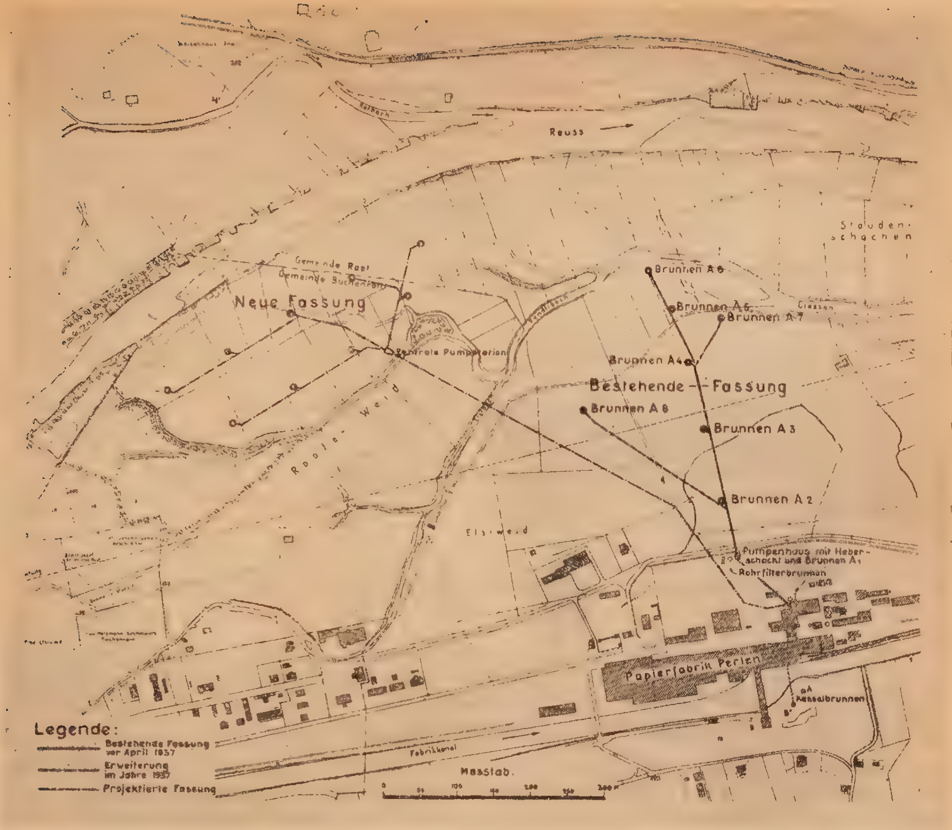


Abb. 1. — Situation der bestehenden und projektierten Fassung im Grundwassergebiet von Perlen.

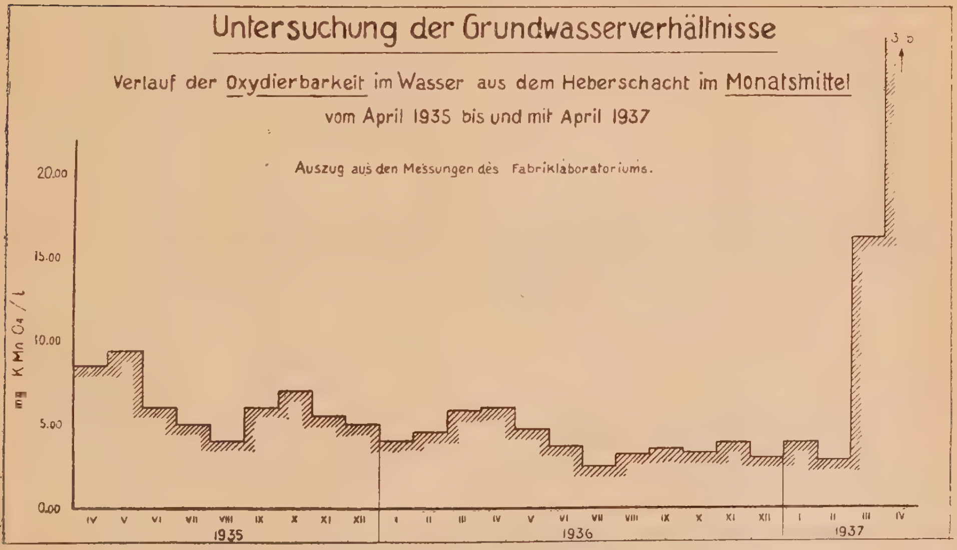


Abb. 2. — Oxydierbarkeit des Grundwassers.

laufe dieser zwei Jahre langsam zurückgingen, erfolgte im April 1937 ein neues, rasches Ansteigen des Kaliumpermanganat-Verbrauches. Im April 1937 wurden Werte bis zu 100 mg/l $KMnO_4$-Verbrauch bestimmt. Dieses für die Fabrikation katastrophale Ansteigen der Verschmutzung im Betriebswasser bedingte umgehende und durchgreifende Massnahmen, durch welche die endgültige Beseitigung dieser lästigen, zu Betriebsstörungen führenden, Verschmutzung gewährleistet wurde.

Die zur Lösung dieser Aufgabe notwendigen Grundlagen mussten zuerst durch Untersuchungen des Fabrikswassers und der zu seiner Fassung dienenden Brunnenanlage auf einer viel breiteren Basis als bisher geschaffen werden. Hiefür sind die hydraulischen Untersuchungen durch das Personal meines Ingenieur-Bureaus in Zürich durchgeführt worden. Die chemischen Untersuchungen waren der Beratungsstelle für Trinkwasserversorgung und Abwasserreinigung an der Eidg. Techn. Hochschule in Zürich zum Teil in Zusammenarbeit mit dem Laboratorium der Papierfabrik Perlen, übertragen worden. Zur Abklärung der geologisch-hydrologischen Fragen wurde noch der bekannte Geologe Dr. J. Hum aus Zürich zugezogen. Im Einvernehmen mit der Fabrikleitung wurde von diesen drei Instanzen folgendes generelle Arbeitsprogramm aufgestellt :

1. Möglichst eindeutige Feststellung der Beschaffenheit der im Brauchwasser periodisch und unregelmässig auftretenden Verschmutzungs-Erscheinungen.
2. Vorschlag derjenigen Massnahmen, welche zu ergreiffen sind, um ein Wiederauftreten dieser Verschmutzungen des Betriebswassers zu vermeiden.
3. Untersuchung des Grundwasserträgers von Perlen auf die Möglichkeit an einer anderen Stelle zusätsliches Wasser zu erschliessen.

B. Untersuchung der bestehenden Fassung und des aus ihr bezogenen Wassers.

Die allgemeine Ausbildung des bis zum Jahre 1934 erstellten Fassungsanlage geht aus Abbildung 3 hervor, einem Schnitt durch die auch im Situationsplan (Abb. 1) eingetragenen fünf Brunnen mit dazugehöriger Heberleitung. Brunnen A_1 ist hiebei als Heber- und Pumpenschacht ausgebildet.

Die Wasseruntersuchungen mussten sich auf jedes der einzelnen fünf verschiedenen Fassungsorgane getrennt erstrecken und es wurden zur abklärenden Bestimmung der Qualität des Wassers und der in ihm enthaltenen Schmutzstoffe die umfangreichen Untersuchungen durchgeführt, wie sie aus den Resultaten vom 12. April 1937 hervorgehen.

Eine zweite ebenso gründliche Untersuchung ist, bei ungefähr auf die Hälfte ihrer Fördermenge gedrosselten Pumpen, am 19. April durchgeführt worden. Ein Teil der anlässlich dieser beiden Untersuchungen erhaltenen Resultate ist in nachtstehender Tabelle aufgeführt :

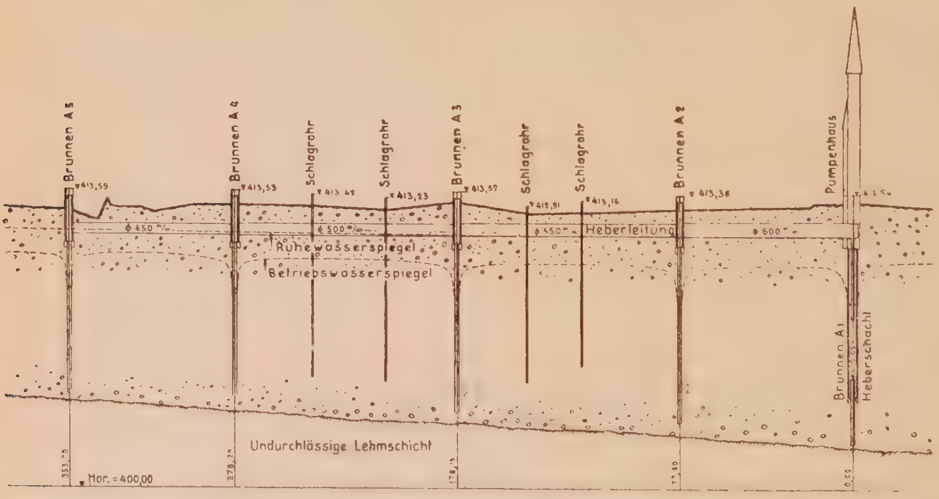


Abb. 3. — Schnitt durch die bestehende Grundwasserfassung der Fabrik mit den Rohrbrunnen A_1, A_2, A_3, A_4, A_5.

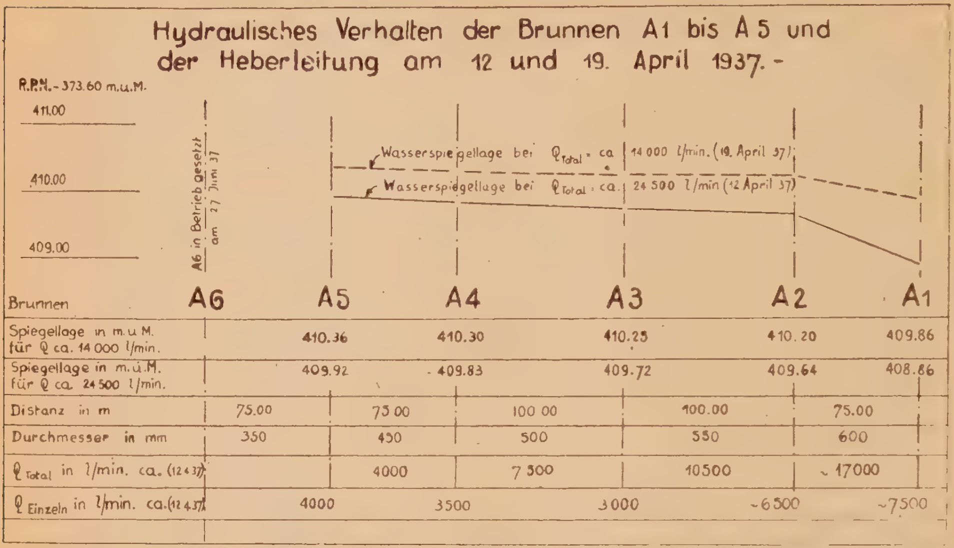
ANALYSEN DES WASSERS AUS BRUNNEN A_1-A_5 IM APRIL 1937

(ausgefürht durch das chem. Labor. der Beratungsstelle für Trinkwasserversorgung an der E.T.H.)

Brunnen	A_1		A_2		A_3		A_4		A_5	
Datum	12	19	12	19	12	19	12	19	12	19
Probe-Entnahme	0.50	3.00	5.00	3.00	5.00	3.00	5.00	3.00	5.00	3.00
Sinnesprufung :										
a) Wasser										
Aussehen	klar	opalesc.	trüb	klar	trüb	klar	trüb	klar	L.trüb	klar
Trübung	gering	gering	stark	keine	stark	keine	flock.	keine	flock.	keine
Farbe	braun	grau	braun	keine	keine	keine	keine	keine	keine	keine
Geruch	erdig	keiner	erdig	keiner	schwach	keiner	keiner	keiner	keiner	keiner
Geschmack	torfig	tintig	tintig	torfig	tintig	bitter	erdig	keiner	keiner	keiner
b) Bodensatz										
Aussehen	Brocken	Flockig	Brocken	Faserig	sandig	sandig	sandig	sandig	sandig	sandig
Farbe	braun	braun	braun	braun	rötl. br.	rötl.	rostig	rot	rötl. grau	rot
Menge in mg /L.	gering	gering	63	gering	88	gering	22	gering	4	gering
Physikal. Untersuchung :										
Temp. in ° C.	9.3	9.1	10.2	10.0	9.3	9.7	7.5	7.5	8.2	8.0
pH - Wert	7.7	7.65	7.75	7.65	7.8	7.7	7.7	7.65	7.7	7.6
Chem. Untersuchung :										
Karbonathärte frz. H°	19.25	18.50	17.75	18.00	16.25	15.75	18.50	16.25	19.50	18.00
Gesamthärte »	28.30	23.60	27.80	23.10	19.85	19.60	23.50	20.00	25.00	20.50
Nichtkarbonathärte »	9.05	5.10	10.05	5.10	3.60	3.85	5.00	3.75	5.50	2.50
aggressive Kohlensäure	0	—	0	—	0	—	0	—	0	—
Trockenrückstand : (105° C) mg /L.	260	220	260	210	270	170	230	175	250	205
Glühverlust : (600° C) Ca CO_3 regen. mg /L.	60	45	35	75	35	45	30	40	30	50
gelöstes Eisen : mg Fe /L	0.15	0.25	0.10	0.10	0.30	0.05	0.03	0.06	0.01	0.07
Gesamteisen im Trockenrückstand : (unfiltriert) mg Fe /L	—	0.30	12.00	0.125	60.00	0.10	2.00	0.25	0.20	0.15
Mangan - Jon : mg /L	0.05	Spur	Spur	Spur		0.00	0.00	0.00	0.00	0.00
Ammoniak : mg NH_3 /L	0.25	0.10	0.05	0.06	0.06	0.08	0.05	0.05	0.06	0.05
Nitrate : mg NO_3 /L	1.47	0.90	1.47	1.30	0.67	0.47	0.67	0.40	3.30	0.70
Nitrite : mg NO_2 /L	0	0	0	0	0	0	0	0	0	0
Sulfate : mg SO_4 /L	26.5	30.0	14.8	20.0	14.6	14.5	14.6	15.1	14.8	14.3
Oxydierbarkeit : mg $KMnO_4$ /L	56.0	34.7	39.0	24.3	12.6	2.2	1.3	1.6	1.3	1.6
Sauerstoffgehalt										
mg O_2 /L	3.83	4.85	5.72	3.17	6.25	6.1	5.94	6.54	4.62	6.9
in % der Sättigung	34.0	42.0	51.0	27.0	55.0	54.0	50.0	49.0	39.0	50.0
Sauerstoffdefizit mg O_2 /L	7.5	6.45	5.5	8.5	5.0	5.2	5.2	6.9	6.6	6.9
Sauerstoffzehrung mg O_2 /L in 48 Std.	0.6	0.67	0.3	0.09	1.8	0.82	0.8	0.2	0.6	1.65

Die Ergiebigkeitsverhältnisse der einzelnen Rohrbrunnen, sowie der Verlauf der Energielinien in der dazugehörenden Heberleitung bei Vollast und bei teilweise gedrosselten Pumpen, geht aus Abbildung 4 hervor. Aus diesen ersten Untersuchungen ergeben sich die folgenden wesentlichen Resultate :

1. Die Qualität des Wassers aus den einzelnen Brunnen ist ausserordentlich verschieden.



Brunnen	A6		A5		A4		A3		A2		A1
Spiegellage in m.u.M. für Q ca. 14 000 l/min.			410.36		410.30		410.25		410.20		409.86
Spiegellage in m.ü.M. für Q ca. 24 500 l/min.			409.92		409.83		409.72		409.64		408.86
Distanz in m		75.00		75.00		100.00		100.00		75.00	
Durchmesser in mm		350		450		500		550		600	
Q_{Total} in l/min. ca. (12.4.37)				4000		7 500		10500		~ 17000	
$Q_{Einzeln}$ in l/min. ca. (12.4.37)			4000		3500		3000		~6500		~7500

Abb. 4. — Ergiebigkeit der einzelnen Rohrbrunnen A_1, A_2, A_3, A_4, A_5.

Das wasser aus den Brunnen A_1 und A_2 ist stark mit eisen und manganhaltigen Stoffen angereichert.

Das Wasser aus Brunnen A_3 enrhält einen hohen Gehalt an suspendierten rostähnlichen Stoffen im Gemisch mit Sand, weist aber im übrigen schon eine viel bessere Zusammensetzung auf.

Das Wasser aus den Brunnen A_4 und A_5 ist ein ausgeglichenes durchmineralisiertes Grundwasser ohne besondere Merkmale.

2. Die Intensität der Verschmutzungserscheinungen ist in starken Masse von der Belastung der Brunnenfassungen abhängig. So ergeben die bei ungefähr halber Vollast der Pumpen durchgeführten Wasseruntersuchungen eindeutig, dass :

a) die Härte, der Gesamteisengehalt, der Mangangehalt und die Oxydierbarkeit stark zurückgegangen sind, und

b) der Sauerstoffgehalt dagegen eine Steigerung von nahezu 100% erfahren hat.

C. Massnahmen zur Behebung der Verschmutzung :

Da aus den Resultaten der chemischen Untersuchung die Zunahme der Verschmutzung mit steigender Beanspruchung der Brunnen bewiesen werden konnte, ergab sich die Möglichkeit der Entlastung der bestehenden Fassung durch den Bau weiterer Rohrbrunnen. Diese mussten aber so plaziert werden, dass sie in Zonen möglichst reinen Grundwassers zu liegen kommen, eine genügende Ergiebigkeit aufweisen, die bestehenden Fassungen in ihrem Ertrag möglichst wenig beeinträchtigen und doch in der Nähe der bestehenden Heberleitung liegen, um lange, kostspielige Zuleitungen zu vermeiden. Um eine in chemischer und hydrologischer Beziehung günstige Zone des Grundwasserträgers zu finden, wurde vorerst ein System von 24 Schlagrohren über das ganze in Frage kommende Gebiet verteilt. Die Lage dieser einzelnen Schlagrohre ist in Abbildung 5 dargestellt. Die aus 2″ Rohren bestehenden,

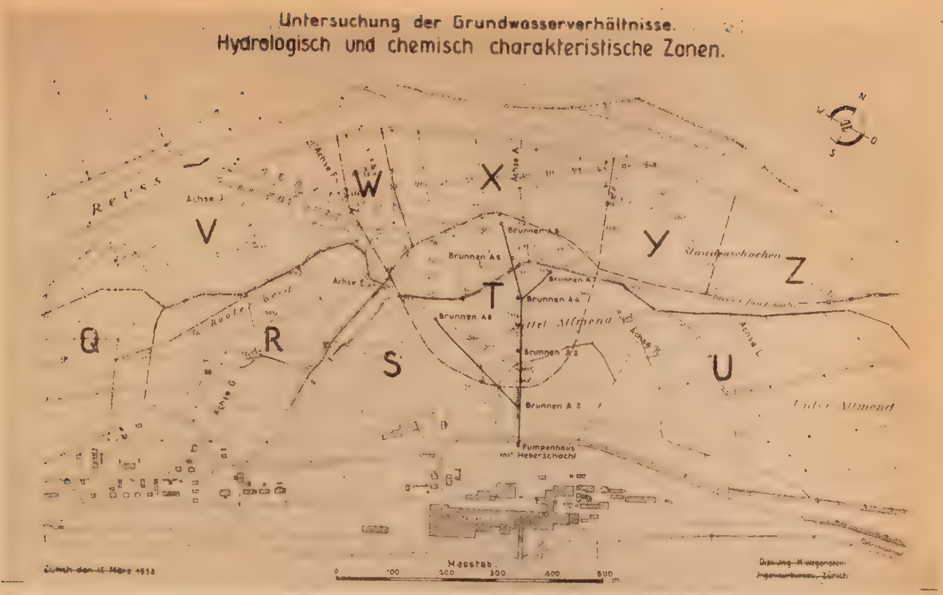


Abb. 5. — Situation aller 98 Schlagrohre und Zoneneinteilung.

in den untersten 2 m gelochten und unten mit einer Eisenspitze versehenen Schlagrohre sind in die Tiefe von 6 - 8 m unter Boden gerammt. Vor Entnahme der Wasserprobe ist jedes einzelne Schlagrohr, mittels einer Hand-Membran-Pumpe, während einer halben Stunde gründlich durchgespült worden.

Die chemische Untersuchung der aus jedem der 24 Schlagrohre zahlreich entnommenen Wasserproben wurde in ähnlicher Weise wie bei der Untersuchung der Wasserproben aus den Brunnen A_1-A_5 durchgeführt. Auf Grund der hiebei erhaltenen Resultate konnte die bestehende Fassung durch Erbohrung und den Anschluss von drei neuen Brunnen A_6, A_7 and A_8 erfolgreich erweitert werden. Diese drei neuen Brunnen mit ihren entsprechenden Anschluss-Heberleitungen sind in Abbildung 1 punktiert eingetragen. Die Detail-Konstruktion eines dieser Brunnen geht aus Ausbildung 6 hirvor, in welcher auch ein für den Grundwasserträger von Perlen typisches Bohrprofil zur Darstellung gebracht ist. Ein Vergleich der spezifischen Ergiebigkeit der einzelnen Brunnen in Funktion der jeweiligen Spiegelabsenkung, wie er in Abbilding 7 graphisch dargestellt ist, lässt die, mit Bezug auf Durchlässigkeit, gleichmässige Beschaffenheit des Grundwasserträgers von Perlen deutlich erkennen. Nach den durchgeführten chemischen Untersuchungen des Wassers aus dem Gebiet der Brunnen A_1 und A_2 ist dasselbe für Trink- und Fabrikationzwecke ungeeignet, sodass diese zwei Brunnen A_1 und A_2 von einer weiteren Benutzung ausgeschlossen wurden. Hiezu wurde in den als Heber- und Pumpenschacht gebauten Brunnen A_1 eine unter Wasser gegossene Betonsohle eingezogen. Brunnen No. 2 konnte einfach durch Schliessen eines Schiebers in der Anschlussleitung abgeschaltet werden.

D. Bau einer neuen Fassung :

Vermittelst der Untersuchungen des Grundwasserträgers van Perlen auf die Möglichkeit an einer andern Stelle zusätzliches Wasser zu erschliessen, musste versucht werden eine Zone zu finden, in welcher :

a) qualitativ absolut sauberes Wasser entnommen werden kann,

b) eine neue Fassung, die im Betrieb bestehenden Brunnen A_3-A_8 nicht oder nur unwesentlich in ihrer Ergiebigkeit beeinträchtigt,

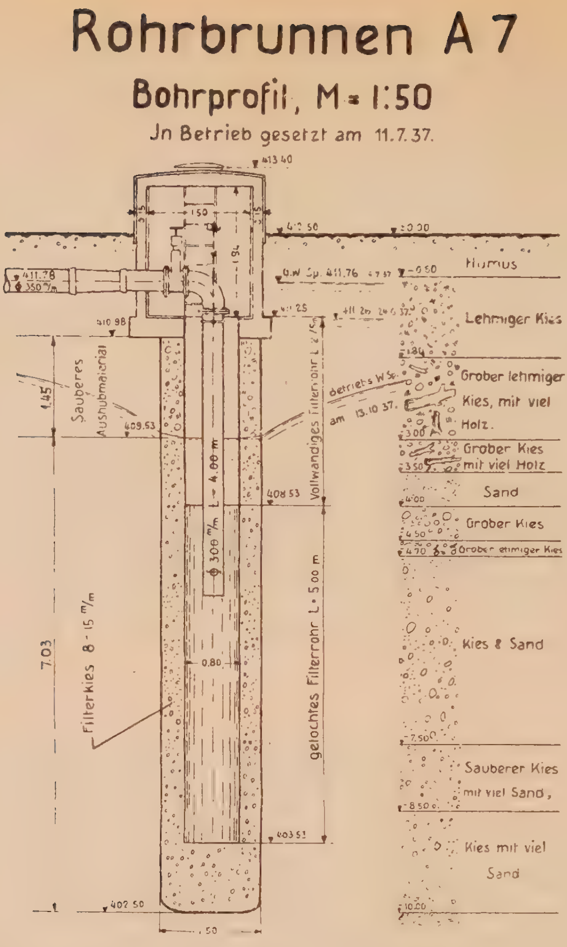


ABB. 6. — Detailplan des Brunnens A_7.

c) der Grundwasserträger eine möglichst grosse spezifische Durchlässigkeit besitzt,

d) und welche in nicht allzu grosser Distanz von den Fabrikgebäuden liegt, um lange und kostspielige Zuleitungen zu vermeiden.

1. *Aufsuchen einer Zone möglichst sauberen Grundwassers* :

Der Grundwasserträger zwischen Fabrik und Fendlibach-Giessen wird durch die Brunnen A_3 - A_8 der bestehenden Fassung weitgehend beansprucht, sodass sich die Untersuchungen für die Erschliessung von neuem Grundwasser auf eine weitere Umgebung erstrecken mussten. Das erste System der 24 Schlagrohre ist daher durch weitere 74 Schlagrohre zwischen Fendlibach-Giessen und Reuss erweitert worden (Abb. 5). Aus allen 98 Schlagrohren sind verschiedene Wasserproben chemisch und biologisch untersucht worden. Die Zahl der von der Beratungsstelle der Eidg. Techn. Hochschule durchgeführten einzelnen Untersuchungen beträgt ca. 1.500. Ein grosser Teil dieser Untersuchungen musste ihrer Natur entsprechend feldmässig, an Ort und Stelle erledigt werden. Ein solches Feldlaboratorium zeigt Abbildung 8 in voller Tätigkeit. Als wesentliches Resultat ergab sich die Erkenntnis, dass das Grundwas-

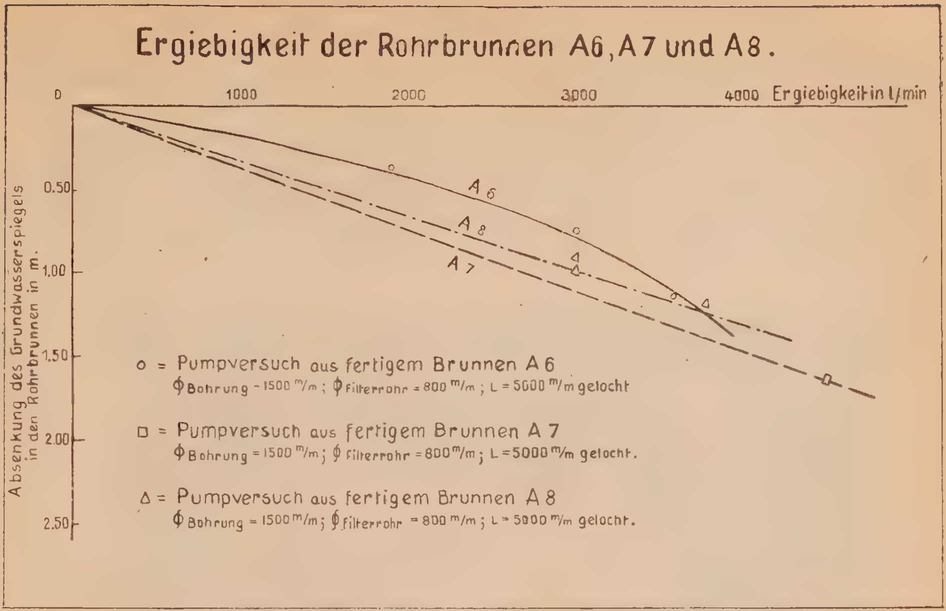


Abb. 7. — Hydrologische Ergiebigkeit der 3 neuen Brunnen A_6, A_7, A_8.

serfeld zwischen der Fabrik und der Reuss keinen einheitlichen kiesigen Grundwasserträger darstellt, sondern dass es sich um ein von überdeckten alten Reuss-Flussläufen durchzogenes, mit Linsen lehmiger, sandiger und torfiger Natur durchsetztes Gebiet handeln muss. Die Qualität des Grundwassers ist von diesen Einlagerungen verschiedenster geologischer Provenienz und dem jeweiligen Grundwasserstand sehr stark abhängig, sodass es unmöglich erscheint, trotz des grossen Anzahl von Einzeluntersuchungen eine irgendwie geordnete oder planmässige Uebersicht zu erhalten.

2. *Untersuchung der Beeinflussung der bestehenden Fassung durch neue Wasserentnahme :*

Hiezu was er notwendig in erster Linie die Beeinflussung der bestehenden Fassung auf den natürlichen, unabgesenkten Grundwasserspiegel festzustellen. Vermittelst Abstichmessungen in den Brunnen der bestehenden Fassung und den 98 Schlagrohren wurde ein Kurvenplan des Grundwasserhorizontes mit einer Aequidistanz von 10 cm aufgenommen und zwar erstmals am 13. Oktober 1937, bei voller Beanspruchung der Brunnen A_3, A_4, A_5, A_6 und A_7 (Abb. 9). Dann wurde versucht ein Kurvenplan des ungestörten Grundwasserspiegels zu erhalten, indem die Pumpen auf ca. 1 /10 ihrer normalen Fördermenge gedrosselt wurden (ein gänzliches Stillegen war aus betriebstechnischen Gründen nicht möglich). Die Verhältnisse bei dem derart stark reduzierten Betrieb und somit praktisch ungestörten Grundwasserspiegel sind aus Abbildung 10 ersichtlich. Ein Vergleich dieser beiden Kurvenpläne zeigt deutlich dass bei der normalen Wasserentnahme aus den Brunnen A_3 - A_7, in der Grössenordnung von ca. 25.000 l /min., das Wasser, ausser der direkten Umgebung, in der Hauptsache den Gebieten zwischen Fassungsanlage und Reuss entnommen wird, wobei sich diese Zone der intensiven Wasserentnahme bis etwa 500 m stromabwärts erstreckt. Beweis hiefür sind die stark abgedrehten Höhenkurven des Grundwasserspiegels im Gebiet des « Staudenschachen », während der Messung vom 13. Oktober 1937 (Abb. 9).

Dagegen wirkt sich die Beeinflussung des natürlichen Grundwasserspiegels stromaufwärts bedeutend weniger weit aus und schon der Verlauf der Niveau-Kurven im Gebiet der Rooter Weid zeigt keine wesentliche Aenderung mehr.

3. *Feststellung der Durchlässigkeits-Verhältnisse* :

Zur Abklärung der Mächtigkeit und der spezifischen Durchlässigkeit des Grundwasserträgers zwischen Fendlibach-Giessen und der Reuss sind in ca 75 m Distanz vom rechten Reussufer, die Probebrunnen SB_{11}, SB_{18}, SB_{21}, SB_{27} abgetäuft worden. Im Verlaufe dieser hydrologischen Untersuchung zeigte sich dann die Notwendigkeit, noch einen weiteren Probebrunnen SB_{14a} zu erstellen. Die Lage aller fünf Brunnen geht aus Abbildung 5 hervor. Aus jedem dieser Probebrunnen wurde ein Pumpversuch in der mittleren Dauer von ca. 160 Stunden durchgeführt, wobei zur Ermittlung der Thiem'schen Absenkungskurve auf zwei zueinander senkrechten Achsen je vier Schlagrohre in Abständen von 3, 15, 35 und 75 m vom Probebrunnen plaziert wurden.

Abb. 8. — Chemisches Feldlaboratorium.

Auf diese Weise konnten die Durchlässigleits-Koeffizienten k für den Grundwasserträger am Ort der fünf verschiedenen Probebrunnen ermittelt werden. Ich verwendete hiezu die aus dem von Darcy-Dupuit aufgestellten Fliessgestezte abgeleitete Formel :

$$k = \frac{Q}{\pi} \frac{ln\, x_2 - ln\, x_1}{Z_2^2 - Z_1^2}$$

wobei
Q Entnahmemenge in in m^3/sek.,
ln natürlicher Logarithmus
x_1, x_2, Abstände der beiden Beobachtungen von der Brunnenachse
Z_1, Z_2, Abstand zwischen den beobachten Grundwasserspiegeln und der undurchlässigen Sohle, bedeutet.

Für die einzelnen Probebrunnen ergaben sich die nachfolgenden, mittleren Ergiebigkeitswerte :

Probebrunnen	Mittlerer k-Wert b. d. einzelnen Probebrunnen	Mittlere k-Werte der Gebiete « Staudenschachen » und « Rooter Weid »
SB 27	0,0076	« Staudenschachen » k = 0,0105
SB 21	0,0134	
SB 18	0,0203	« Rooter Weid » k = 0,0213
SB 14a	0,0186	
SB 11	0,0250	

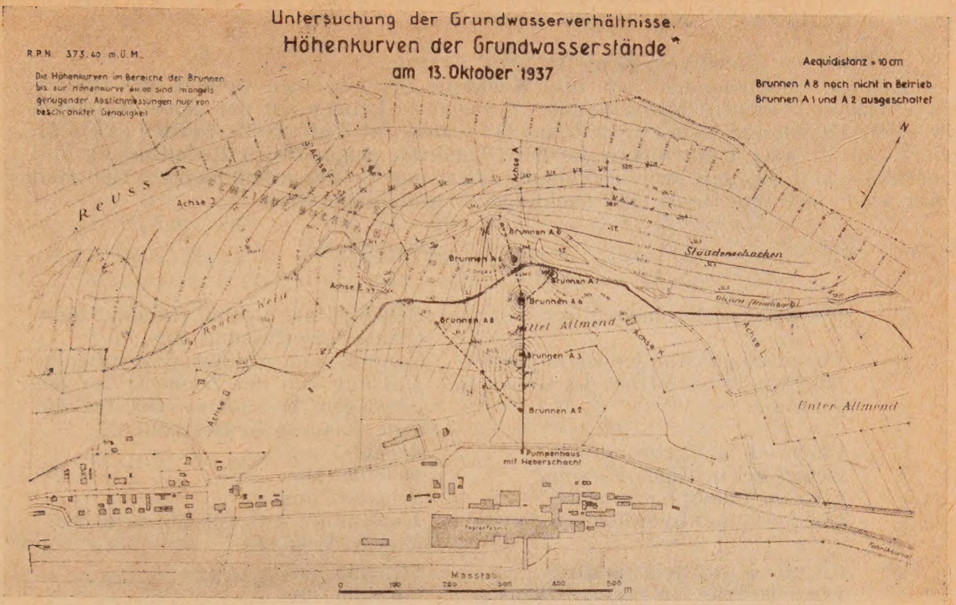

Untersuchung der Grundwasserverhältnisse.
Höhenkurven der Grundwasserstände*
am 13. Oktober 1937
R.P.N. 373.60 m.ü.M.
Die Höhenkurven im Bereiche der Brunnen bis zur Höhenkurve ... sind mangels genügender Abstichmessungen nur von beschränkter Genauigkeit
Aequidistanz = 10 cm
Brunnen A 8 noch nicht in Betrieb
Brunnen A 1 und A 2 ausgeschaltet
N
Reuss
Achse 3
Achse A
Achse E
Achse G
Achse K
Achse L
Brunnen A 5
Brunnen A 4
Brunnen A 7
Brunnen A 6
Brunnen A 8
Brunnen A 3
Brunnen A 2
Staudenschachen
Mittel Allmend
Unter Allmend
Pumpenhaus mit Heberschacht
Masstab
0 100 200 300 400 500 m

ABB. 9. — Grundwasserhorizont bei voller Beanspruchung der bestehenden Fassung.

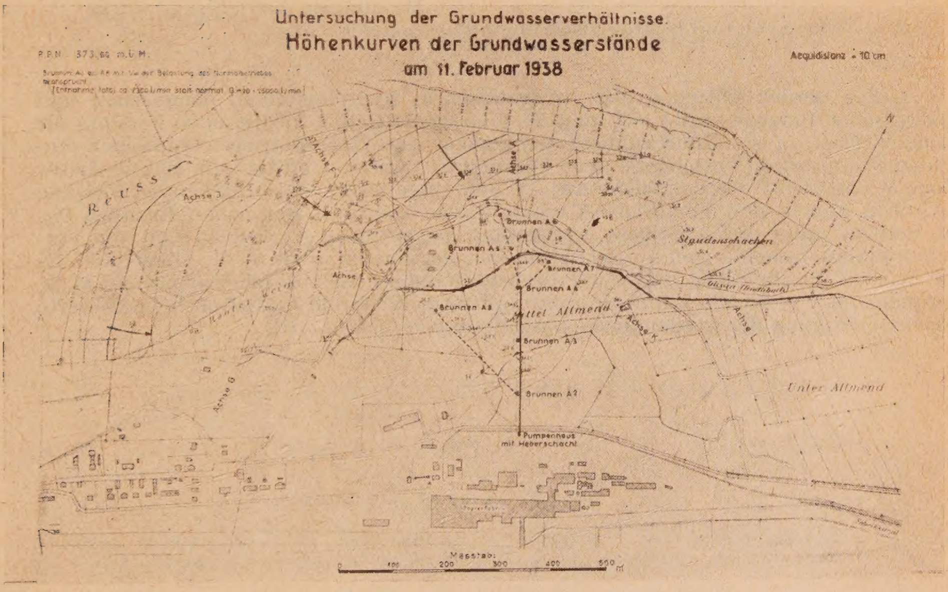


ABB. 10. — Grundwasserhorizont bei stark reduzierter Beanspruchung der bestehenden Fassung.

Hieraus ergeben sich eindeutig die mit Bezug auf Durchlässigkeit günstigeren Verhältnisse im Gebiet der « Rooter Weid ».

4. *Wahl der günstigsten Zone für neue Fassung* :
Um die verschiedenen Vor- und Nachteile der einzelnen Gebiete des Grundwasserträgers eindeutig gegeneinander abwägen zu können, ist der ganze Talboden zwischen Fabrik and Reuss in 10 Zonen Q bis und mit Z eingeteilt worden (Abb. 5). Die Resultate aller Untersuchungen betr. Qualität, gegenseitiger Beeinflussing und Durchlassigkeit aus den für eine Fassung überhaupt in Frage kommenden Gebieten zwischen Fabrik und Reuss sind nachfolgend zusammengestellt :

Zone	Befund	Beurteilung
Q	Chemisch nicht einwandfrei	Scheidet als Gebiet für eine neue Grundwasserfassung aus.
R	Chemisch günstig, jedoch in demselbenGrundwasserstrom wie Gebiet Q gelegen	Kommt nach Abklärung der Beeinflussung von der Zonne Q her bei günstigem Befund als Ort für die neue Fassung in Betracht.
S	Chemisch nicht einwandfrei	Scheidet aus
T	Gebiet der heutigen Fassung	Scheidet aus
U	Chemisch nicht einwandfrei	Scheidet aus
V	Chemisch und hydrologisch günstig	Kommt als Ort für die neue Fassung in Betracht
W	Chemisch nicht einwandfrei	Scheidet aus
X	Einzugsgebiet der bestehenden Fassung	Scheidet aus
Y	Chemisch und hydrologisch günstig	Da diese Zone zur Enthname der geforderten Betriebswassermenge von 20.000 l/min. zu klein ist, könnte sie höchstens als Ort einer Teilfassung in Frage kommen. Es dürfte möglich sein, aus diesem Gebiet eine Wassermenge von 6.000 8,000 l/min. zu entnehmen, dessen Qualität nach den vorliegenden Messungen befriedigen würde.
Z	Chemisch und hydrologisch ungünstig	Scheidet aus

Unter Berücksichtigung aller hydraulischen, hydrologischen, chemischen und geologischen Belange ergibt sich nun als günstigstes Gebiet für die neue Fassung die Zone V. Diese, aus 9 Rohrbrunnen, den zugehörigen Heberleitungen und einer zentralen Pumpstation bestehende neue Fassung ist daher endgültig zum Bau in dieser Zone V vorgesehen worden (siehe Abb.1). Seit Erweiterung der bestehenden Fassung zwischen Fabrik und Fendlibach-Giessen durch Anschluss der neuen Brunnen A_6, A_7 und A_8, unter gleichzeitigem Ausschalten der Brunnen A_1 und A_2 sind die Verschmutzungserscheinungen im Betriebswasser der Fabrik nicht mehr aufgetreten. Es ist daher wahrscheinlich, dass auch die neue Fassung das zusätzlich benötigte Wasserquantum von 20.000 l/min. in einwandfreiem Zustande und ohne nennenswerte Beeinträchtigung der vorhandenen Fassungsanlage liefern wird.

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